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The Pb–Pb Trail Dating of Carbonates with Open U–Pb Systems: The Min'yar Formation of the Upper Riphean Stratotype, Southern Urals

G. V. Ovchinnikova*, I. M. Vasil'eva*, M. A. Semikhatov, I. M. Gorokhov*, A. B. Kuznetsov*, B. M. Gorokhovskii*, and L. K. Levskii***

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Abstract—Epigenetic alterations and secondary Pb–Zn mineralization in carbonate rocks of the Min'yar Formation represent a serious problem for the direct measurements of Pb–Pb ages in these upper horizons of the Upper Riphean stratotype (the Karatau Group of southern Urals). In addition to petrographic data, cathodoluminescent analysis, and strict geochemical criteria (Mn/Sr, Fe/Sr, $\delta^{18}O$), we paid attention, when selecting samples for Pb–Pb dating, to the minimum difference between $^{87}Sr/^{86}Sr$ ratios measured in secondary carbonate phases leached in the 1 N ammonium acetate solution and in primary phases residual after leaching. In order to remove secondary carbonates and microinclusions of Pb–Zn sulfides as carefully as possible, we subjected the preliminary selected samples to a short-term leaching in 0.5 N solution of HBr before their final dissolution in this acid. The Pb–Pb dates estimated for whole rock samples and for residual carbonate phases after the last leaching are equal to 780 ± 200 and 780 ± 110 Ma, respectively. The Pb–Pb date of 470 ± 80 Ma obtained for four carbonate phases (whole rock carbonates, two their fractions leached in 1 N NH_4OAc and 0.5 N HBr solutions, and the residual phase) of one sample bearing sulfide microinclusions is considered as the age of Pb–Zn mineralization. The model age t_1 = 770 ± 220 Ma is calculated for dolomites on the basis of "isochron t_1–t_2" assuming that t_2 = 470 Ma. The weighted average age value for the Min'yar dolomites is 780 ± 85 Ma. This result well fits the formation age limits evaluated on the basis of direct data of isotopic geochronology, paleontology, and chemostratigraphy. The tested procedure of sample selection and isotopic study of carbonate rocks opens a possibility to measure Pb–Pb ages of stratigraphic importance in those cases, when U–Pb systems of rocks were reset in the course of epigenetic sulfide mineralization.

Key words: **U–Pb systematics, carbonate rocks, epigenesis, Upper Riphean, southern Urals.**

INTRODUCTION

Isotopic–geochronological characteristics of the Upper Riphean stratotype (Karatau Group of the western slope of the Urals) are known so far from results of K–Ar and Rb–Sr dating of glauconite inadequately studied in mineralogical aspect and from K–Ar dates obtained for whole rock samples of gabbro-diabase. At the same time, the recent progress in U–Pb and Pb–Pb dating methods offered a possibility to determine age of many Precambrian carbonate-bearing successions (Moorbath *et al.*, 1987; Jahn *et al.*, 1990; Jahn and Cuvellier, 1994; Jahn and Simonson, 1995; Ovchinnikova *et al.*, 1995; Russell *et al.*, 1996). In particular, the Pb–Pb isochron age for limestones from the middle part of the Karatau Group (Inzer Formation) is determined as equal to 836 ± 25 Ma (Ovchinnikova *et al.*, 1998b). This value is interpreted as indicating age of the early diagenesis in carbonate sediments, and the date obtained is consistent with Rb–Sr dates (805–835 Ma) measured in authigenic illite fractions originated in the same formation during the burial diagenesis (Gorokhov *et al.*, 1995).

In this work, we analyzed the U–Pb and Pb–Pb systems of carbonate rock samples from the Min'yar Formation that overlies the Inzer Formation and characterizes an upper part of the Karatau Group. In this case, the Pb–Pb dating procedure meets some problems owing to epigenetic Pb–Zn sulfides dispersed in the Min'yar rocks (Mirkina *et al.*, 1986; Shirobokova, 1992). The mineralizing fluid was able to considerably decrease U/Pb ratios and to change the Pb isotope composition within a large volume of carbonate deposits, where the initial Pb concentration is extremely low, <1 μg/g (Ovchinnikova *et al.*, 1998a). Accordingly, we had to elucidate influence of secondary Pb–Zn mineralization on the U–Pb and Pb–Pb systems of the Min'yar dolomites, to demonstrate the applicability of U–Pb dating for carbonate rocks subjected to such alteration, and to estimate age of the formation.

STRATIGRAPHIC POSITION, AGE, AND COMPOSITION OF STUDIED SAMPLES

The Karatau Group of siliciclastic–carbonate deposits (3.0–5.5 km thick) is exposed within the Bashkirian meganticlinorium in the western slope of southern Urals. In the regional stratigraphic scheme accepted now, the group is divided into six units (from the base upward): Sil'merdak (1200–3300 m), Katav (200–350 m), Inzer (300–800 m), Min'yar (350–800 m), Uk (160–350 m), and Krivaya Luka (250–300 m) formations. The lower and upper units are siliciclastic, whereas others are composed, predominantly or entirely, of carbonate deposits (Kozlov *et al.*, 1991; Maslov and Krupenin, 1991; Semikhatov *et al.*, 1991; *Unifitsirovannye...*, 1993; Maslov *et al.*, 1998).

Composition, succession, and chronological relationships of almost all formations of the Karatau Group have been established quite a long ago (see *Stratotip rifeya...*, 1983), but the structural and lithostratigraphic interpretation of boundary beds separating the Min'yar and underlying Inzer formations remained ambiguous for a while (see review in Maslov and Krupenin, 1991). In the west of the Bashkirian meganticlinorium (Adzhigardak Ridge), where we studied the Min'yar succession (Fig. 1), the formation in question is situated between the upper Inzer and lower Uk siliciclastic members. The succession that is bounded at the base and crowned at the top by thin limestone members consists mainly of monotonous 350–380 m thick dolomits (Fig. 2) enclosing, within the upper portion, the abundant nodules, lentils, and interlayers of chert (Kozlov *et al.*, 1991).

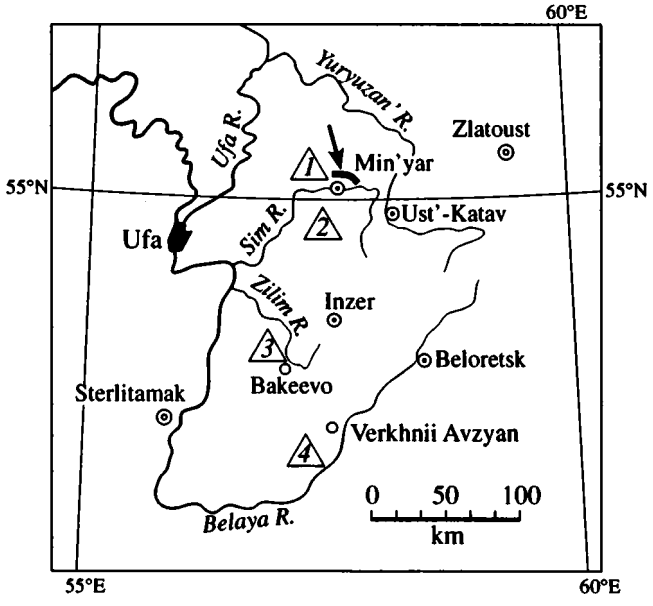

Fig. 1. Geographic position of the studied Min'yar section and of the Ivanovsk (*1*), Srednyaya Kurgashla (*2*), Zilim (*3*), and Pravoberezhnoe (*4*) sites of Pb–Zn mineralization in the western slope of southern Urals.

During a rather long period, it was *a priori* accepted that the Karatau Group represents almost completely the stratigraphic range of the Upper Riphean (e.g., Keller and Semikhatov, 1968; *Stratotip rifeya...*, 1982, 1983; Keller *et al.*, 1984; Semikhatov *et al.*, 1991; Koroteev *et al.*, 1997). This opinion was favored by K–Ar dates obtained in the 1960s–1970s for globular glauconites from different levels of the group, although this material was inadequately studied in the mineralogical aspect. These dates depict a regular, decreasing upward succession of age values: 938 Ma for the upper part of the Katav formation, 896–853 Ma for the lower and middle parts of the Inzer Formation, 791–740 Ma for the uppermost member of the latter, 713–680 Ma for the Min'yar Formation, and 658–630 Ma for the basal member of the Uk Formation (Garris, 1977; *Stratotip rifeya...*, 1983 and references therein). The Rb–Sr isochron age measured for the latter is 688 ± 10 Ma (Gorozhanin and Kutyavin, 1986). The upper age limit of the Karatau Group (*Stratotip rifeya...*, 1983; Semikhatov *et al.*, 1991; Koroteev *et al.*, 1997) was inferred from several evidences: K–Ar dates for whole rock samples of gabbro–diabase crosscutting the succession (660–620 Ma); K–Ar dates for glauconites from basal Vendian beds unconformably overlying the group in the Urals and nearby region (625–560 Ma); and from U–Pb dates for zircons from granitoids intruding the presumable analogues of the Karatau Group westward of its distribution area (660 ± 15 Ma), though these dates obtained by A.A. Krasnobaev have never been validated in publications. The estimated lower age limit of the group is based on extrapolation of K–Ar dates for glauconites from the lower Karatau horizons and on occasional K–Ar whole rock dates for gabbro-diabase bodies, which presumably originated before deposition of the latter and yielded, in addition to distinctly rejuvenated results, age values of about 1150–1000 Ma (Garris, 1977).

The listed dates, which seem to be unreliable now in methodical aspect (especially those obtained for the whole rock samples), have been considered off and on as arguments favoring the idea that the Karatau Group represents sufficiently well the succession of rocks ranging in age from 1000 ± 50 to 650 ± 20 Ma. We should mention here, however, that the significance and time range of stratigraphic hiatus obviously separating the Karatau Group from overlying Lower (tillites of the Kurgashlya Formation) or Upper Vendian deposits (clastic rocks of the Asha Group) have never been considered, and erosion marks at the base of the Min'yar and Uk formations (Bekker, 1961, 1988; Raaben, 1975; Kozlov, 1982; Kozlov *et al.*, 1991; Maslov and Krupenin, 1991) were either ignored without comments, or considered as evidencing the synsedimentary growth of local uplifts.

Leaving here the problem of lower age limit of Karatau deposits, we would like to emphasize that paleontological remains from the middle–upper interval of the succession are characteristic of relatively high lev-

els of the Upper Riphean. The Min'yar Formation encloses remains of *Melanocyrillium* (Maslov *et al.*, 1994), and these vase-shaped protistans appeared in geological records about 850 Ma ago (Knoll, 1996). In addition, stromatolites from the Min'yar and Uk formations (Raaben, 1975) and organic-walled microfossils from the lower Inzer strata (*Stratotip rifeya...*, 1982) are similar to their counterparts occurring in the upper Upper Riphean of Spitsbergen and some other regions (Raaben, 1975; Butterfield *et al.*, 1994).

Problems concerning the stratigraphic range and age interpretation of upper subdivisions of the Karatau group showed their unexpected sides in the light of recent chemostratigraphic considerations based on secular variations of carbon (Semikhatov *et al.*, 1995; Podkovyrov *et al.*, 1998) and Sr isotope ratios (Gorokhov *et al.*, 1996a, 1996b; Kuznetsov *et al.*, 1997, 1998a).

First, it was demonstrated that considerable intervals with high $\delta^{13}C_{carb}$ values (from +6 to +8–9‰PDB), which characterize the successions ranging in age from 850 to 630 Ma in many other regions and are thought to represent an immanent feature of C-isotope variations recorded at that time (Kaufman and Knoll, 1995; Bartley *et al.*, in press), are missing from the Min'yar and Uk successions. In opinion of Podkovyrov, Semikhatov and their colleagues (Podkovyrov *et al.*, 1998), this phenomenon just reflects considerable gaps in the upper Karatau carbonate succession that is conditioned by accumulation of siliciclastic deposits (Krivaya Luka Formation) before the pre-Vendian erosion and by hiatuses in sedimentation at the base of Min'yar and Uk formations, the accumulation time of which was relatively short. For instance, the accumulation of the latter is interpreted as a short-term event that immediately followed the Sturtian glaciation about 700 Ma ago.

Second, C and Sr isotopic characteristics of Karatau carbonates (*op cit.*) suggest that the so-called Shaler event, i.e., a sudden decrease of $^{87}Sr/^{86}Sr$ initial ratios recorded in the synonymous supergroup of Arctic Canada (Asmerom *et al.*, 1991), corresponds in the Urals to the late-Min'yar changes in chemostratigraphic parameters. The Shaler Supergroup appears to be younger than 1100 Ma (the U–Pb age of youngest population of clastic zircons from its base) and older than 723 ± 3 Ma (the U–Pb age of baddeleyite from volcanics discordantly overlying the Supergroup; Heaman and Rainbird, 1990; Heaman *et al.*, 1992; Rainbird *et al.*, 1992). The telecorrelation suggests that this supergroup could be even younger than 802 ± 10 Ma (Rainbird *et al.*, 1996). Accordingly, it seems reasonable to connect the Shaler event with the Rodinia supercontinent breakup that commenced 750–725 Ma ago (Powel *et al.*, 1993; Torsvik *et al.*, 1996). In other words, if all chemostratigraphic considerations are correct, the Min'yar Formation is older than 750–725 Ma, whereas paleontological data cited above show that it is younger than 850 Ma.

We collected the studied samples of carbonate rocks from the Min'yar Formation stratotype exposed near and inside the synonymous town (Figs. 1 and 2). Dominant here are light gray to pinkish massive or thin-platy stromatolitic and mechanogenic dolomites intercalated with flakestone lentils. In the upper portion of the succession, they enclose abundant nodules and lentils of dark chert, the first occurrence level of which is commonly accepted for the boundary between the lower (200–210 m) and upper (160–190 m) subformations. The basal and topmost members (10–15 m) are composed of limestone; the former incorporates a packet of siliciclastic rocks (1–3 m).

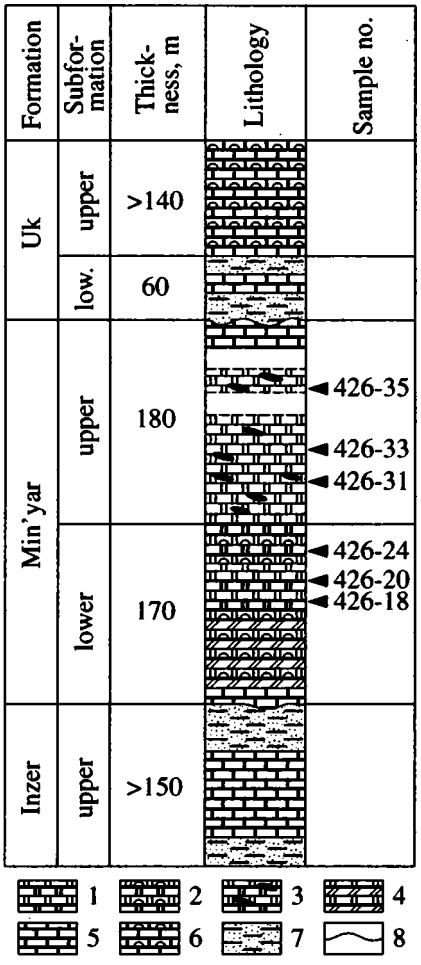

Fig. 2. Lithology and labeled stratigraphic levels of studied samples in the Min'yar Formation section: (1) dolomite; (2) stromatolitic dolomite; (3) limestone; (4) dolomite with cherty inclusions; (5) low-Mg dolomite; (6) stromatolitic limestone; (7) siliciclastic deposits; (8) stratigraphic unconformity.

The Min'yar carbonate rocks have the insoluble silicate residue (3% in average) consisting of quartz and less abundant feldspar with a negligible admixture of illite and chlorite (Gareev, 1986). Stromatolitic rock varieties correspond to dolomicrites and dolo(micro)sparites, whereas microsparites and fine-grained sparites are of mechanogenic origin. Small veinlets and inclusions of late crystalline dolomite, abundance of which is irregu-

lar, are characteristic of all rock types. The rocks preserve original bedding structures and clastic texture, though they are noticeably recrystallized according to data of cathodoluminescent analysis. The reddish luminescence visible by analysis varies from weak to moderate, being most intense in the late dolomite segregation. In zonal crystals of that dolomite present in dolosparites, the core and marginal zones differ in the luminescence intensity (Podkovyrov *et al.*, 1998).

In the upper subformation, cherty nodules and lentils are composed of microcrystalline quartz locally associated with chalcedony. Silicified microfossils, some perfectly preserved, are dispersed in these cherty intercalations (Sergeev, 1992). The occurrence of microfossils suggests that cherts of the Min'yar Formation, like many other Upper Precambrian cherts, originated immediately after the postmortal burial of microorganisms in the surface sedimentary layer near the water–sediment interface (Maliva *et al.*, 1989). At the same time, cherty nodules and lentils of the Min'yar Formation locally enclose the dolomicrosparite relicts, and some flakestones distinctly show that silicification predated synsedimentary brecciation of dolomicrosparites. Consequently, the early dolomitization of Min'yar carbonates took place at the initial stage of lithogenesis immediately after accumulation of sediments composed presumably of high-Mg calcite. This does not concerns, however, the late dolomite generation filling veins and local nests in the rocks. This generation is most characteristic of low-Mg dolomites localized in the lower part of the formation, where the rocks experienced the late dedolomitization (Kuznetsov *et al.*, 1998b).

Carbonate sediments of the Min'yar Formation accumulated in a vast marine basin that occupied, in addition to the western slope area of the Urals, also a very wide zone nearby. Structural features of the rocks suggest a transition from relatively deep sedimentation settings at the early Min'yar time to the shallow settings located near the fair weather wave base in the subsequent period (Maslov, 1997).

The Pb–Zn sulfide mineralization in carbonate rocks of the Karatau Group is concentrated mostly in southern and eastern areas of the Bashkirian meganticlinorium, but in the west, where we studied the Min'yar section, it is manifested to a much lesser extent (Mirkina *et al.*, 1986; Shirobokova, 1992). Ore occurrences proper are situated here as far away from the town of Min'yar as 20–30 km (see Fig. 1). In the studied succession, thin (0.2–2.0 mm) veins and segregations of small galena and sphalerite (less abundant) crystals tend to be concentrated in horizons of coarse-grained dolomite, which are localized near the contact of carbonate and siliciclastic deposits.

ANALYTICAL PROCEDURE

Selection of carbonate samples most appropriate for the U–Pb isotope analysis included two consequtive steps. First, we selected visually those 17 of 38 available dolomite samples, which looked homogeneous and affected by secondary recrystallization to a minimum extent. One half of each sample was used for the X-ray, chemical, cathodoluminescent, and isotopic analyses, and another for investigation under petrographic and electron microscopes. Element concentrations in carbonate rock fractions were analyzed by chemical (Ca and Mg) and atomic absorption (Fe and Mn) methods after dissolving weightings of samples in 1 N HCl. We did not use more aggressive acids in order to leave intact the silicate fraction.

Afterward, we selected 6 dolomite samples fitting the following criteria: $Mn/Sr < 1.5$; $Fe/Sr < 4.0$; $Rb/Sr < 0.2$; $\delta^{18}O > -5‰$ PDB. As is known (Brand and Veizer, 1980, 1981; Veizer, 1983; Banner and Hanson, 1990; Asmerom *et al.*, 1991; Derry *et al.*, 1992; Gorokhov *et al.*, 1995; Gorokhov, 1996; Kuznetsov *et al.*, 1997; and references in these works), the above criteria characterize "best" dolomite samples that suffered to a minimum extent the epigenetic (diagenetic in English literature) alteration by interaction with meteoric waters and are, consequently, most appropriate for further investigation.

Precambrian carbonates usually include non-cogenetic phases with different Sr isotope composition and trace element concentrations. Only one of the phases is of the early diagenetic origin, whereas all others are products of secondary (catagenetic) alterations (Gorokhov *et al.*, 1995; Ovchinnikova *et al.*, 1995, 1998a, 1998b). In order to separate these phases, we treated the powdered samples in 1 N solution of ammonium acetate (NH_4OAc) under the room temperature (Gorokhov *et al.*, 1995; Ovchinnikova *et al.*, 1995). The carbonate residue after this treatment representing the primary rock component is designated here as PCC-1 phase (Fig. 3). It is important that the non-cogenetic carbonate phases of selected samples show the minimum difference, in terms of Rb–Sr systematics, between the measured $^{87}Sr/^{86}Sr$ ratios, and, consequently, the degree of secondary alterations in relevant rocks is at the minimum. The whole rock carbonate fraction used without treatment in NH_4OAc solution is designated as PCC-2 phase. To extract U and Pb from both phases, we dissolved them in 1 N HCl under the room temperature. The residue insoluble in HCl represents the ISR phase, and the mixture of concentrated HF and HNO_3 was used to decompose it.

Since carbonate rocks of the Karatau Group are affected by the epigenetic Pb–Zn mineralization, we had to use the additional investigation and treatment methods by analyzing six selected samples. Three of them (426-18, 426-24, and 426-35) have been studied under the scanning electron microscope ABT-55 equipped with microprobe LINK AN-1000. Small

pyrite crystals up to 6 × 12 μm have been detected only in sample 426-24 (Fig. 4). However, it is impossible to guarantee, even after examination under scanning electron microscope, that other samples are free of tiny sulfide crystals less than 6 μm in size that is comparable with the microprobe resolution ability. Accordingly, we used the hard leaching method in order to remove not only possible sulfide microinclusions, but also the carbonate phases formed or recrystallized by the development of Pb–Zn mineralization.

We should emphasize here that application of the Pb–Pb isochron model and differentiated dissolution of carbonate fractions, is a promising approach as only in this case we can obtain a reliable age information analyzing only composition of daughter Pb isotopes without measuring concentrations of parental uranium. It is obvious that dissolution procedure is unable to change the Pb isotope composition bearing geochronological information in any rock phase under analysis. Changes in U/Pb ratios in response to differentiated dissolution are unimportant for the Pb–Pb age calculations.

Reagents commonly used for hard leaching of Precambrian rocks are either 0.1 M and 6 M solutions of HCl (Whiterhouse and Russell, 1997), or 0.6 N and 1.2 N solutions of HBr (Babinski *et al.*, 1995). In our study, we used the 0.5 N solution of HBr (Vasil'eva *et al.*, 1998) bearing in mind the following criteria: (1) sulfide solubility is higher in HBr than in HCl; (2) final dissolution of carbonate phases in the less concentrated 0.5 N HBr is more preferable than in the 6 N HCl (or even in 1 N HCl), because a chance to extract Pb from siliciclastic admixture decreases; (3) it is a common practice to extract Pb for mass-spectrum analysis filtering the 0.5 N bromide solution through the ion-exchange columns, and, having no necessity to convert chloride into bromide, we decrease a possibility of laboratory contamination.

Fractions of carbonate rock sample (PCC-1 or PCC-2) were treated during 30 minutes in 0.5 N HBr under the room temperature. The insoluble residue was rinsed three times with water after centrifuging, dried up, and weighted. About 30% of initially weighed portion entered the solution. This dissolved material is designated as L-1 phase (Fig. 3). Carbonate fraction of dry residue after the 30-minute-long leaching was treated once again during 24 hours in 0.5 N HBr to be completely dissolved. This fraction is designated as L-2 phase. In all phases, we determined U and Pb concentrations and analyzed the Pb isotope composition. Subjected to the hard leaching are PCC-1 phases of samples 426-18, 426-24, and 426-33, and also PCC-2 phases of samples 426-20, 426-31, and 426-35, that is why indices L-1 and L-2 are supplemented in the table with abbreviations pointing to one or another phase. Likewise, ISR phases residual after carbonate dissolution in different acids (HCl or HBr) are respectively marked in the table. Because of a low Pb concentration in dolomites, the relatively large batches of samples

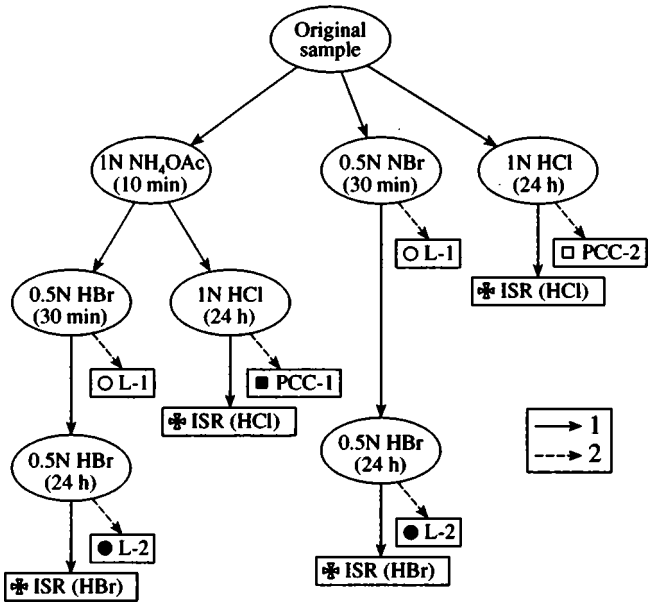

Fig. 3. Scheme of chemical treatment of studied samples and separated insoluble (ISR) and soluble (PCC) phases: (1) insoluble residue after sample treatment in the indicated reagent; (2) soluble phase (centrifugate).

(600–800 mg) were used for the two-step leaching in HBr. Accordingly, data on PCC-1 and PCC-2, on the one hand, and on their residues L-1 and L-2 are obtained from different sample batches.

The mixed tracer $^{235}U + {}^{208}Pb$ was added to aliquots of bromide solutions used to analyze the U and Pb concentrations and the Pb isotope composition. Extraction of U and Pb in bromide forms was performed using the anion-exchange resin Bio-Rad 1x8 in accordance with procedure described by (Manhes *et al.*, 1978). U and Pb isotope compositions were analyzed using mass spectrometer Finnigan MAT-261 in a regime of simultaneous measurements of all isotope ion currents. The measured Pb isotope ratios were corrected for a fractionation factor of 0.13% per atomic mass init based on recurrent determinations of Pb isotope composition in NBS SRM-982 standard. The procedure blank during the period of measurements is 0.05 ng for U and 0.5 ng for Pb. The computation algorithm by Ludwig (PBDAT) was used to estimate uncertainties and correlation of U–Pb data, and his ISOPLOT program was used to calculate the isochron parameters and MSWD values (Ludwig, 1989, 1990). All inaccuracy values presented in the text and Fig. 6 correspond to two standard deviations (2σ).

RESULTS

The least altered samples selected to study the U–Pb systematics of the Min'yar Formation characterize its middle strata (Fig. 2) that are situated away from both the basal low-Mg epigenetic dolomites and the recrys-

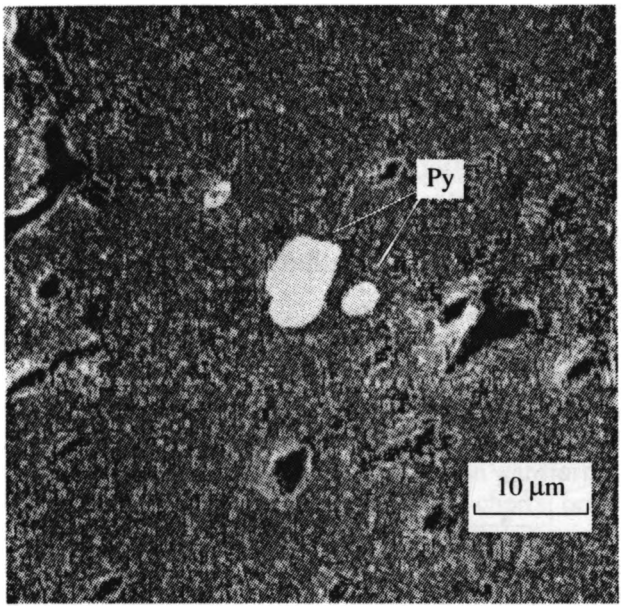

Fig. 4. SEM photograph of pyrite microinclusions (Py) in dolomite sample 426-24 from the Min'yar Formation.

tallized rock varieties crowning the exposed interval of the studied section. As a result of burial and meteoric diagenetic events, dolomites of the last type show elevated values of Mn/Sr (1.4–16.4) and Fe/Sr (4.1–64.5) ratios. As compared to others, the "best" dolomite samples from the middle part of the section mostly consist of pure dolomite (Mg/Ca = 0.606–0.636) and have an insignificant non-carbonate admixture (0.7–2.6%) and micrite relicts, but they are almost free of the late, fine- to medium-grained dolomite generation. The Mn, Fe, and Rb contents in these samples are low in contrast to a rather high Sr concentration (Table 1), and consequently, Mn/Sr and Fe/Sr ratios characterizing the relevant rocks are low (0.61–1.53 and 1.46–3.55, respectively). Values of $\delta^{18}O$ vary in the rocks from –4.0 to –4.8‰ PDB, and difference between $^{87}Sr/^{86}Sr$ ratios measured in PCC-1 and PCC-2 phases (one subjected to and another left without treatment in ammonium acetate) is less than 0.0025. The listed characteristics prove that the studied rocks did not experience a considerable alteration at the burial and meteoric diagenetic stages (Choquette and James, 1987; Barnaby and Read, 1992; Montanez and Read, 1992; Gorokhov *et al.*, 1995; Gorokhov, 1996; Kuznetsov *et al.*, 1998b).

Data on U–Pb systems in different carbonate fractions of the samples studied are shown in Table 2. According to these data, the "delicate" leaching reveals the following characteristics: (1) U and Pb concentrations are higher in the PCC-1 phases of samples 426-20, 426-31, and 426-35, than in their PCC-2 phases, and the former simultaneously have lower values of measured $^{238}U/^{204}Pb$ ratios and less radiogenic Pb isotope composition; (2) in contrast, PCC-1 phases of samples 426-18 and 426-33 show $^{238}U/^{204}Pb$, $^{206}Pb/^{204}Pb$, and $^{207}Pb/^{204}Pb$ ratios that are higher than those in PCC-2 phases; (3) all three isotopic parameters are almost identical (within the limits of analytical error) in PCC-1, PCC-2, and L-1 phases of sample 426-24, and only the $^{206}Pb/^{204}Pb$ ratio in L-2 phase is in this case higher by 6% than in others. The established relationships between parameters suggest that leaching in NH_4OAc partially removes the epigenetic carbonate phases (or admixture), where Pb may be relatively more (samples 426-20, 426-31, and 426-35) or less radiogenic (samples 426-18 and 426-33). At the same time, the extraction of secondary carbonate phases from sample 426-24 does not change parameters of its U–Pb isotopic system.

In addition, Table 2 shows that the highest values of $^{238}U/^{204}Pb$, $^{206}Pb/^{204}Pb$, and $^{207}Pb/^{204}Pb$ ratios characterize the L-2 phases of samples 426-18, 426-24, 426-31, and 426-33, which represent residues after leaching in 0.5 N HBr. However, in samples 426-20 and 426-33 enriched in radiogenic Pb, the highest values of these parameters are typical of PCC-2 and PCC-1 phases, and this may indicate that carbonate phases of these particular samples are heterogeneous.

Data points corresponding to PCC-1 and PCC-2 phases commonly useable for age determinations produce a linear trend in the $^{207}Pb/^{204}Pb$–$^{206}Pb/^{204}Pb$ diagram of Gerling and Houtermans (Fig. 5). The Pb–Pb age calculated using the linear trend of PCC-1 data points is equal to 988 ± 400 Ma, MSWD = 22, and that for PCC-2 phases corresponds to 775 ± 130 Ma,

Table 1. Chemical composition of carbonate rock fraction in dolomite samples from the Min'yar Formation

Sample no.	Insoluble residue, %	Ca, %	Mg, %	Mn, μg/g	Fe, μg/g	Rb, μg/g	Sr, μg/g	Mg/Ca	Mn/Sr	Fe/Sr	$\delta^{18}O$, ‰ PDB
426-18	1.8	21.7	13.20	44	143	0.12	54.9	0.608	0.80	2.60	–4.0
426-20	2.6	22.1	13.38	101	145	0.14	74.7	0.606	1.35	1.94	–4.3
426-24	0.9	21.8	13.36	51	100	0.15	68.5	0.612	0.74	1.46	–4.8
426-31	1.6	21.6	13.22	68	90	0.49	44.4	0.612	1.53	1.98	–4.6
426-33	0.7	20.9	13.28	63	150	0.18	103	0.636	0.61	1.46	–4.3
426-35	2.2	22.0	13.41	96	240	0.13	67.6	0.609	1.42	3.55	–4.4

Table 2. U–Pb data for different phases of carbonate rocks from the Min'yar Formation

Sample no.	Phase	Pb*, μg/g	U*, μg/g	$^{238}U/^{204}Pb$	$^{206}Pb/^{204}Pb$	$^{207}Pb/^{204}Pb$	$^{208}Pb/^{204}Pb$
426-18	PCC-2	0.271	0.126	34.5	28.863	16.401	39.770
	PCC-1	0.227	0.127	43.2	31.537	16.603	40.321
	L-1 (PCC-1)	0.278	0.166	31.1	29.011	16.395	39.612
	L-2 (PCC-1)	0.194	0.176	71.5	33.047	16.710	40.649
426-20	PCK-2	0.244	0.217	80.7	44.479	17.428	42.312
	PCK-1	0.365	0.254	56.3	35.265	16.797	40.749
	L-1 (PCC-2)	0.285	0.197	59.6	40.144	17.073	41.469
	L-2 (PCC-2)	0.306	0.253	69.8	37.907	17.034	41.490
	ISR (HBr)	0.552	1.689	216.1	26.182	16.185	38.439
426-24	PCC-2	1.014	0.160	10.6	22.393	15.993	38.253
	PCC-1	0.939	0.159	11.4	22.592	15.998	38.312
	L-1 (PCC-1)	0.956	0.156	10.9	22.298	15.940	38.240
	L-2 (PCC-1)	0.653	0.282	29.5	23.740	16.066	38.447
426-31	PCC-2	0.366	0.207	43.8	33.631	16.683	38.427
	PCC-1	0.514	0.233	33.3	29.534	16.389	38.180
	L-1 (PCC-2)	0.494	0.226	33.6	29.590	16.377	38.115
	L-2 (PCC-2)	0.338	0.234	54.3	34.803	16.810	38.549
	ISR (HCl)	1.600	3.527	14.8	23.516	16.054	37.188
	ISR (HBr)	2.721	5.486	13.5	23.165	16.081	37.423
426-33	PCK-2	0.272	0.142	40.7	31.003	16.569	41.584
	PCK-1	0.211	0.162	64.3	35.795	16.900	43.554
	L-1 (PCC-1)	0.241	0.242	79.3	31.774	16.584	41.841
	L-2 (PCC-1)	0.227	0.145	52.3	33.926	16.787	42.746
426-35	PCC-2	0.530	0.292	39.5	27.949	16.327	37.632
	PCC-1	0.627	0.326	36.7	26.899	16.253	37.715
	L-1 (PCC-2)	0.778	0.321	28.8	25.883	16.162	37.652
	L-2 (PCC-2)	0.489	0.317	46.9	28.929	16.445	37.599
	ISR (HCl)	2.876	6.057	13.6	20.627	15.838	37.254
	ISR (HBr)	2.845	13.77	31.3	20.605	15.826	37.345

* Concentrations of U and Pb in leachates are calculated relative to the mass difference between PCC-2 phases and residues after leaching.

MSWD = 12. The MSWD values obtained indicate that both calculated ages are unreliable in terms of statistics. In the $^{206}Pb/^{204}Pb$–$^{238}U/^{204}Pb$ diagram, dispersion of data points is even greater (Fig. 6), and we can definitely state that the U–Pb systems of studied samples have been reset in the course of their geological history.

DISCUSSION

According to results obtained, even the "best" Min'yar dolomites that passed the preliminary geochemical test do not fit the isochron model conditions in terms of the U–Pb systematics. The necessary conditions of the model are the uniform composition of primordial Pb and the steady-state equilibrium in the Pb–Pb and U–Pb system of rocks during the whole geological history of the latter, whereas the established dispersion of data points in Pb–Pb diagrams may be a consequence of the U loss and/or the Pb contamination.

Carbonate rocks that lost uranium under the recent hypergene conditions have been described in a series of publications (e.g., Asmerom and Jacobsen, 1993; Jahn and Cuvellier, 1994; Hoff *et al.*, 1995; Johnes *et al.*, 1995; Ovchinnikova *et al.*, 1995). However, events of this kind were hardly crucial in the history of U–Pb systems in the Min'yar carbonates, as it is evident from the fact that the $^{238}U/^{204}Pb$ ratios and U concentrations measured in different phases of all samples show a poor correlation (see Table 2). At the same time, the studied dolomite samples reveal the following symptomatic characteristics:

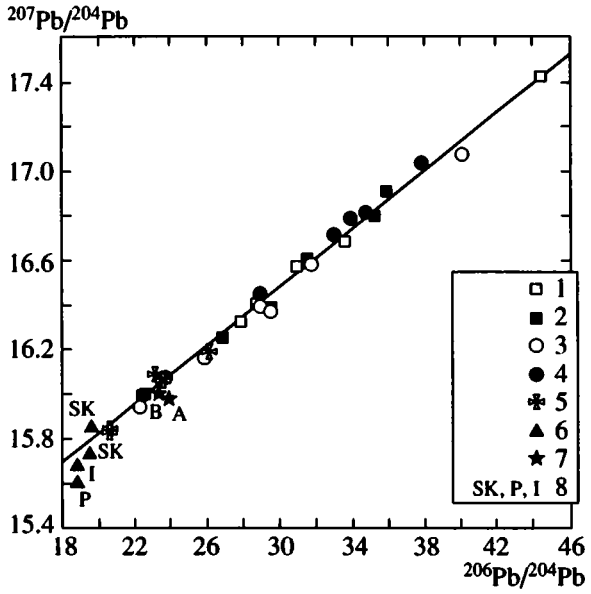

Fig. 5. $^{207}Pb/^{204}Pb$–$^{206}Pb/^{204}Pb$ diagram for different carbonate phases of dolomite samples from the Min'yar Formation and for galena specimens from Pb–Zn mineralization sites: (1–5) carbonate phases PCC-2 (1), PCC-1 (2), L-1 (3), L-2 (4), and ISR (5), see text for explanation; (6) galena from the Srednyaya Kurgashla (SK), Pravoberezhnoe (P), and Ivanovsk (I) sites of ore mineralization (after Mirkina *et al.*, 1986); (7) two calculated isotopic compositions of allogenic Pb.

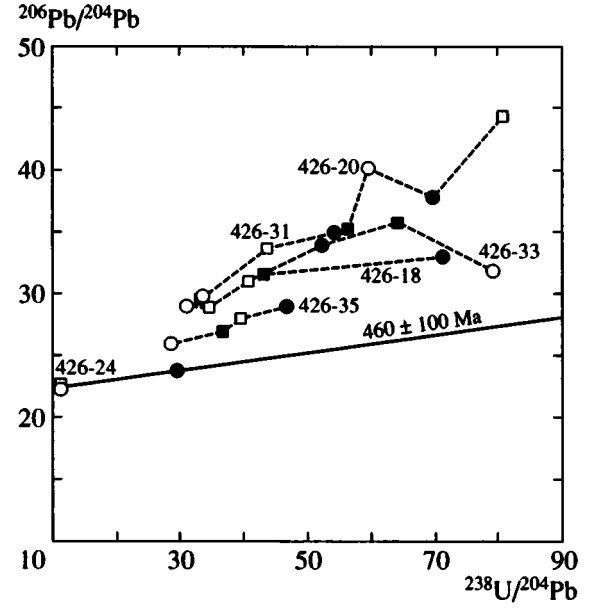

Fig. 6. $^{206}Pb/^{204}Pb$–$^{238}U/^{204}Pb$ diagram for different carbonate phases of dolomite samples from the Min'yar Formation. Dotted lines connect data points characterizing sample, the number of which is indicated nearby. The U–Pb age of 460 ± 100 Ma is calculated using data on PCC-1, PCC-2, L-1, and L-2 phases of Sample 426-24 (other symbols as in Figs. 3 and 5).

(1) Phases of the same type from various samples and different phases of particular sample show the negative correlation between the Pb content and measured $^{238}U/^{204}Pb$ ratios (Fig. 7).

(2) The highest Pb concentration but lowest $^{238}U/^{204}Pb$ ratio are established in carbonate fractions of one sample (426-24), where pyrite microinclusions have been recognized by examination under the scanning electron microscope.

(3) The Pb isotope composition in PCC and L phases from the last sample is approaching that characterizing phases insoluble in HCl (ISR phases in Table 2 and Fig. 5) and also galena from the Pb–Zn ore deposits of the Bashkirian meganticlinorium (Mirkina *et al.*, 1986).

(4) The laboratory leaching in ammonium acetate that is aimed to remove secondary carbonate phases from samples (Gorokhov *et al.*, 1995; Ovchinnikova *et al.*, 1995) does not result in Pb extraction.

(5) Removing secondary carbonates and decomposing sulfides, the hard leaching in 0.5 N HBr noticeably increases Pb concentration in acid leachates (L-1 phases) relative to the residual L-2 phases, which are most depleted in Pb.

All these characteristics suggest that dolomites of the Min'yar succession have been contaminated with Pb at a certain stage of their geological history, and this explains well the data point dispersion in Pb–Pb diagrams (Fig. 5). This allogenic Pb slightly mobile under leaching in NH_4OAc (PCC-1phases) enters solution much easier under leaching in 0.5 N HBr (L-1 phases in Fig. 7). Consequently, dolomite samples from the Min'yar succession yield two types of leads different in isotopic composition and in behavior under leaching. One appears to be fixed in the dolomite crystal lattice (primary carbonate phase), and another one is a component of easily solvable sulfides and, maybe, of epigenetic carbonates that have been formed at the surface of dolomite grains.

The allogenic Pb origin in the Min'yar dolomites can be elucidated by analyzing diagrams of $^{208}Pb/^{204}Pb$ and $^{206}Pb/^{204}Pb$ versus $^{207}Pb/^{204}Pb$. In these diagrams, all data points for carbonate phases, except for those of Sample 426-20, plot along straight lines (Figs. 8 and 9), which can be considered as mixing lines of primary and allogenic leads. The most radiogenic Pb is typical of phases L-2, whereas the least radiogenic Pb characterizes phases L-1. Since the latter contain, in addition to allogenic Pb, a certain share of Pb from carbonates, isotopic parameters of the allogenic Pb should be searched for along the mixing lines to the left of data points representing phases L-1 (Figs. 8 and 9). These parameters can be estimated using a version of the "isochron family model" that was suggested by Tera (1981) and then successfully used by Neimark and Iskanderova (1987), who calculated isotopic parameters of primary Pb in potassium feldspars from Proterozoic granitoids of Baikal Highland.

As is known, marine sedimentary carbonates lack correlation between $^{206}Pb/^{204}Pb$ and $^{208}Pb/^{204}Pb$ ratios characterizing respectively the uranium and thorium decay chains (Jahn and Cuvellier, 1994). Correspondingly, data points for PCC-2 phases in Figs. 8 and 9 reflect variations of Th/U ratios in the mixture of primary carbonate and allogenic Pb. The inclination angles of lines connecting data points of L-1, L-2, PCC-1, and PCC-2 phases in $^{208}Pb/^{204}Pb$–$^{206}Pb/^{204}Pb$ ($^{207}Pb/^{204}Pb$) diagrams depend on the sample ages and Th/U ratios in their different phases. If, for instance, two samples of identical age with dissimilar Th/U ratios contain allogenic Pb of the same isotopic composition (entering by leaching the phase L-1), the lines connecting phases of each sample must have an intersection point with Pb isotope ratios characterizing the allogenic lead. Data points for Sample 426-20 do not reveal a linear correlation in the $^{208}Pb/^{204}Pb$–$^{206}Pb/^{204}Pb$ plot, whereas the inclination angles of other correlation lines show that Th/U ratios in the mixture of primary and secondary materials of samples 426-18 and 426–33 are higher than in that of samples 426–31 and 426–35.

An insignificant difference between Pb isotope ratios in various phases of any sample studied is responsible for a high uncertainty of angle coefficients calculated for the mixing lines, and we can outline only an area of their intersection points. Isotopic parameters of intersection points were calculated for two line pairs, one corresponding to samples 426-31 and 426-35, and another to samples 426-33 and 426-18. These points (A and B, respectively) characterizing the isotopic composition of allogenic Pb are shown in Fig. 5. Note also that data point of residual and solvable carbonate phases of Sample 426-24, as well as those of galena, plot close to the area, where all the mixing lines intersect (Figs. 8 and 9).

The origin of sulfide mineralization in the Karatau deposits has been interpreted differently (Mirkina *et al.*, 1986; Shirobokova, 1992; Seravkin *et al.*, 1994). According to one viewpoint, lead involved into mineralization was leached by epigenetic fluids from siliciclastic and carbonate rocks of the Karatau group (Anfimov, 1984; Mirkina *et al.*, 1986), and this idea is compatible with Pb isotopic parameters obtained for galena from the Min'yar Formation and for the residual ISR phases. In Fig. 5, data points for galena samples produce a trend that intersects the errorchron plotted for carbonate phases, and the ISR phases plot close to the intersection point thus being attributable to this errorchron. In addition, data points characterizing the calculated isotopic composition of allogenic Pb also plot close to the above intersection. Thus, it seems possible that Pb–Zn mineralization was accompanied by Pb leaching out of siliciclastic and carbonate rocks composing the upper formations of the Karatau Group. The mineralizing fluid could be also responsible for the partial recrystallization of carbonate components at the

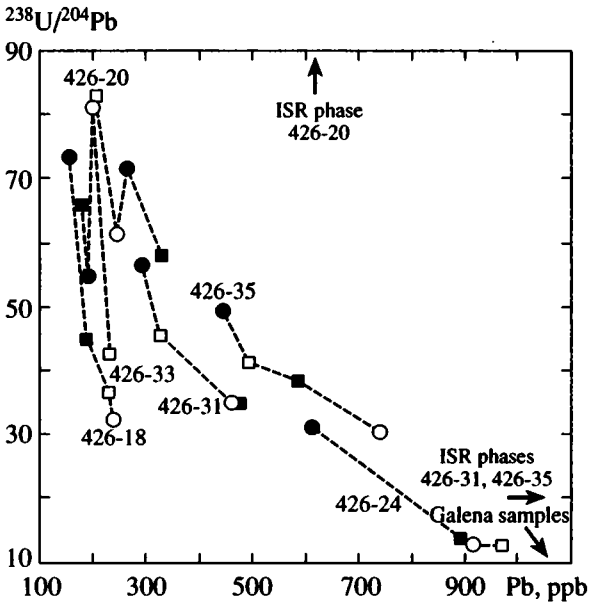

Fig. 7. Plot of $^{238}U/^{204}Pb$ ratios versus Pb concentration in various carbonate phases of indicated dolomite samples from the Min'yar Formation (symbols as in Figs. 3 and 5).

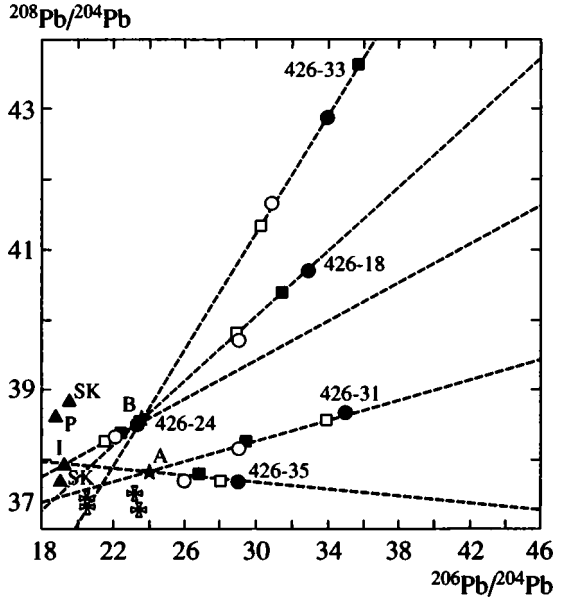

Fig. 8. $^{208}Pb/^{204}Pb$–$^{206}Pb/^{204}Pb$ diagram for different carbonate phases of indicated dolomite samples from the Min'yar Formation (symbols as in Figs. 3 and 5).

time of precipitation of Pb–Zn sulfides. The highest concentration of trapped Pb is detected in Sample 426-24, where it is 2–4 times higher than in other samples, whereas the $^{238}U/^{204}Pb$ ration is 3 to 8 times lower. In this connection, we consider data on Sample 426-24 as inappropriate for the Pb–Pb age calculation.

In the diagram of $^{206}Pb/^{204}Pb$ versus $1/^{204}Pb$ (Fig. 10), data points of PCC-1, PCC-2, L-1 (leachate after sample

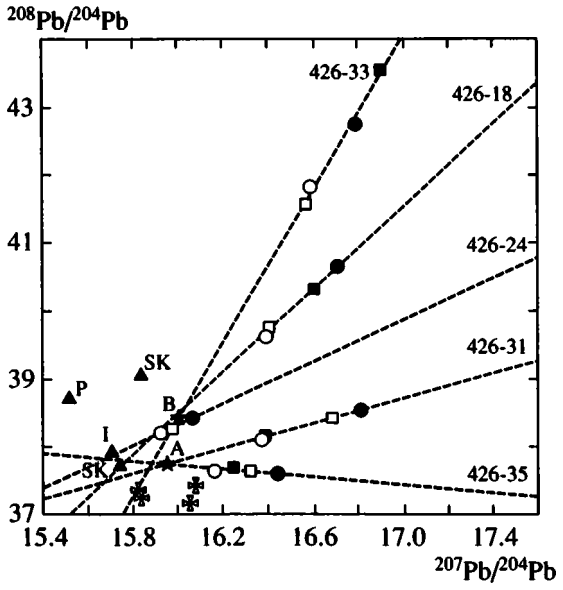

Fig. 9. $^{208}Pb/^{204}Pb$–$^{207}Pb/^{204}Pb$ diagram for different carbonate phases of indicated dolomite samples from the Min'yar Formation (symbols as in Figs. 3 and 5).

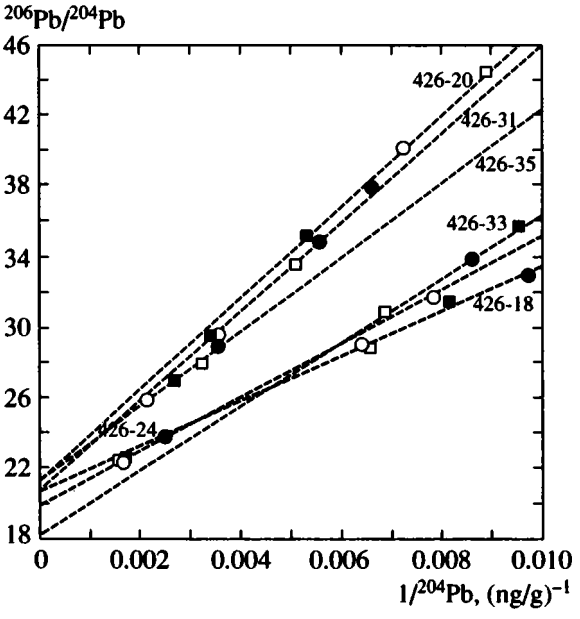

Fig. 10. $^{206}Pb/^{204}Pb$–$1/^{204}Pb$ diagram for different carbonate phases of indicated dolomite samples from the Min'yar Formation (symbols as in Figs. 3 and 5).

treatment in HBr), and residual L-2 phases plot along straight lines. Consequently, we may suspect that each particular sample exemplifies mixing of leads variable in isotopic composition. Note also that one-type phases of all samples do not show any common trend, and this indicates that each sample yields two types of leads, one of the carbonates proper, which consists of primordial and radiogenic components, and another allogenic Pb.

The allogenic Pb of samples 426-33 and 426-18 differs in isotopic composition from that of samples 426-31 and 426-35. Data points A and B characterizing the allogenic lead plot below the Pb–Pb isochron (Fig. 5), close to data points of Sample 426-24, but quite away from those exemplifying PCC-2 phases of all other samples. Hence, we may assume that PCC-2 and L-2 phases of studied samples, except for those of Sample 426-24, have an insignificant concentration of allogenic Pb and can be used for calculating the Pb–Pb age valid in geochronological aspect. For five data points of PCC-2 phases, the calculated value is 780 ± 200 Ma, MSWD = 6.

Lead in L-2 phases residual after leaching in 0.5 N HBr should be pretty close in isotopic composition to the lead of primary carbonates, because the general level of $^{207}Pb/^{204}Pb$, $^{206}Pb/^{204}Pb$, and $^{238}U/^{204}Pb$ ratios is highest exactly in these phases. Age value calculated on the basis of five L-2 phases is 780 ± 110 Ma, MSWD = 1.5, and the last parameter suggests that this date is statistically reliable.

The Pb–Pb isotopic parameters of carbonate rocks bearing Pb–Zn mineralization depend on the time, when mineralizing fluids affected the rocks, and on the degree of secondary recrystallization of carbonates. If sulfides and carbonates are approximately synchronous in origin, then their Pb isotope composition had to be balanced at the early diagenetic stage thus being usable for Pb–Pb dating (Jahn and Cuvellier, 1994; Ovchinnikova *et al.*, 1998a). In this case, the inclination of straight line in the $^{207}Pb/^{204}Pb$–$^{206}Pb/^{204}Pb$ diagram is geochronologically meaningful. The case was illustrated by data on the Schmidtsdrift dolomite succession in the Upper Archean Transvaal Supergroup (South Africa), which is bearing the stratiform sulfide mineralization (Jahn *et al.*, 1990). If the mineralization is superimposed on carbonate rocks formed earlier, we have to consider two possibilities: (1) the allogenic Pb source of unknown isotopic composition must confuse correlation of data points for carbonates in the $^{207}Pb/^{204}Pb$–$^{206}Pb/^{204}Pb$ plot (Winter and Johnson, 1995); (2) the linear approximation of data points has in this case an ambiguous geochronological meaning. In general, the linear correlation of this type represents the mixing line that could be interpreted sometimes as the "isochron t_1–t_2," which offers a chance to estimate age of initial event (t_1), if we know the age of superimposed process (t_2). Relationships of this kind were suggested for the Lower Proterozoic marbles in the Sjangeli foldbelt (northern Sweden and Norway), which became enriched in radiogenic Pb during the Caledonian events. As a result, the straight line approximating data points of marbles in the $^{207}Pb/^{204}Pb$–$^{206}Pb/^{204}Pb$ diagram turned out to be inclined steeper than the Pb–Pb isochron corresponding to formation time t_1 of original carbonate rocks (Romer, 1994). Suggesting that the obtained straight line represents the "isochron t_1–t_2," Romer calculated age t_2 corresponding to recrystallization time of carbonates and demonstrated that the

inferable t_1 value corresponds to the U–Pb age of zircons from basal tonalites of the Sjangeli Complex.

It is unknown how old is the Pb–Zn mineralization in carbonate rocks of the Min'yar Formation. This mineralization of distinct epigenetic origin (Feoktistov *et al.*, 1978; Mirkina *et al.*, 1986; Shirobokova, 1992) seems to be a consequence of concentration of dispersed ore components in response to activation of tectonics and magmatism in the study region. Events of this kind characterize the whole history of the Uralides and pre-Uralides (Ivanov *et al.*, 1986; Puchkov, 1993, 1997; Ivanov, 1998) and have been accompanied from time to time by development of ore mineralization. For instance, the Ordovician deposits in the east of western slope of the Urals, which accumulated on the passive continental margin, enclose the stratiform polymetallic and Ba-polymetallic mineralization related in origin to redistribution of synsedimentary ore components by mineralizing fluids (Shirobokova, 1992; Seravkin *et al.*, 1994).

As is demonstrated above, the correlation between the $^{206}Pb/^{204}Pb$ ratios and Pb concentrations in the studied samples of the Min'yar dolomites characterizes mixing of Pb trapped in carbonates (primordial and radiogenic components) with Pb added to the rocks by epigenetic mineralizing fluids (see Fig. 10). In the $^{207}Pb/^{204}Pb$ – $^{206}Pb/^{204}Pb$ diagram, data points characterizing carbonate phases of samples 426-18, 426-31, 426- 33, and 426-35 (except for the L-2 phase of the latter) plot along the almost parallel straight lines (dotted lines in Fig. 11), which run close to data points representing isotopic parameters of hypothetical allogenic Pb (asterisks in Fig. 11) and those measured in galena (triangles in Fig. 11). Like in the case of carbonates from the Sjangeli Complex (Romer, 1994), these mixing lines slope steeper than the Pb–Pb isochrons approximating the PCC-2 or L-2 phases, the first of which yields age value of 780 Ma (see Fig. 10). Slope ratios of mixing lines correspond to age values of 1030–1190 Ma. If the allogenic Pb was leached from carbonate rocks, whose U–Pb systems developed in a closed state from the early diagenesis time (t_1) to the commencement of recrystallization (t_2), the mentioned mixing lines can be considered as "isochrons t_1–t_2." Then, assuming one age parameter, e.g., t_2, we may calculate another using the equation:

$$\tan\alpha = 1/137.88(\exp\lambda_5 t_1 - \exp\lambda_5 t_2)/(\exp\lambda_8 t_1 - \exp\lambda_8 t_2),$$

where λ_5 and λ_8 are the decay constants of ^{235}U and ^{238}U, respectively.

The t_2 value for carbonates of the Min'yar Formation appears to be concordant to the age of epigenetic sulfide mineralization. According to results of leaching experiments, the Pb isotope ratios in all phases of Sample 426-24 contaminated by sulfides are close to the ratios calculated for the allogenic Pb, and it is likely

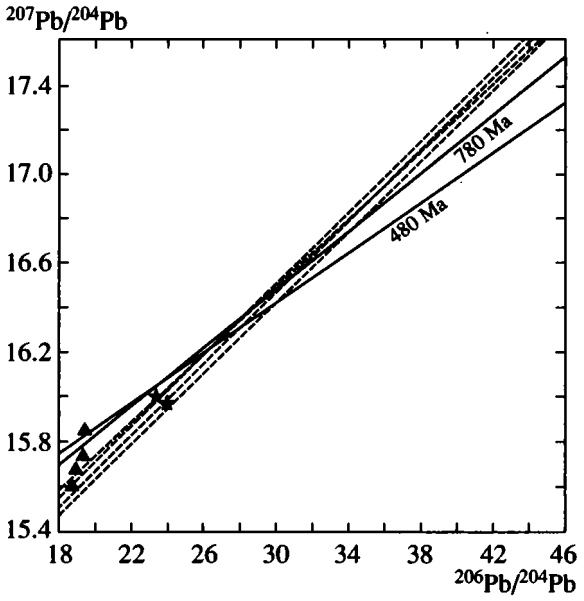

Fig. 11. $^{207}Pb/^{204}Pb$–$^{206}Pb/^{204}Pb$ diagram for dolomite samples from the Min'yar Formation. Dotted lines correspond to "isochrons t_1–t_2" calculated for samples 426-18, 426-31, 426-33, and 426-35 (data points are omitted). Solid lines are isochrons for PCC-2 phases of studied samples (780 Ma) and for carbonate phases of Sample 426-24 (480 Ma). Other symbols correspond to those used in Figs. 3 and 5.

that the Pb–Pb system of this sample was completely reset in the course of Pb–Zn mineralization. As the Pb isotope composition in L-2 phase is more radiogenic than in other carbonate phases of the sample, a certain amount of Pb accumulated here owing to U decay after the time of Pb–Zn mineralization.

The Pb–Pb age calculated on the basis of data points characterizing carbonate phases of Sample 426–24 is equal to 480 ± 100 Ma (see. Fig. 11), and their U–Pb age is estimated to be 460 ± 100 Ma (Fig. 6). Despite a considerable uncertainty of both dates that is related to a narrow variation range of Pb isotopic ratios, one is very close to another, and the average value of 470 ± 80 Ma determines the Early Ordovician age of Pb-Zn mineralization. In the western zone of the Urals, this period corresponded to the initial rifting stage responsible for the crustal shortening in diverging passive margins of the ancient continental masses, whereas in the Uralide zone located eastward of the Main Uralian fault, this was already a time of sea-floor spreading development (Ivanov *et al.*, 1986; Puchkov, 1997; Ivanov, 1998). Geodynamic environments of this kind are commonly thought to be manifestations of the mantle plume activity favorable as well for generation of thermal fluids, which could mobilize ore elements and produce Pb–Zn mineralization in the Karatau Group deposited on the newly formed continental passive margin.

If the U–Pb and Pb–Pb systems of carbonates were reset at the time t_2 = 470 ± 80 Ma, then the t_1 value cal-

culated using the aforementioned equation is 770 ± 200 Ma (an average age value for PCC-2, L-1, and L-2 phases of samples 426-18, 426-31, 426-33, and 426-35). This date is concordant with dates of 780 ± 200 and 780 ± 110 Ma calculated respectively for PCC-2 (carbonates untreated in ammonium acetate) and L-2 phases (residual carbonates after the first-step leaching in 0.5 N HBr).

We should remark here that the model of "isochron t_1–t_2" that was successfully used by interpretation of Pb isotopic data on sulfides and feldspars (Neimark, 1988; Ovchinnikova *et al.*, 1994; Amelin *et al.*, 1998) is correct only then, when it is applied for samples, which are almost free of uranium and radiogenic Pb accumulated after the time t_2. Dolomites of the Min'yar Formation do not fit this requirement, because they accumulated a certain amount of radiogenic Pb after the time of Pb–Zn mineralization. However, the $^{238}U/^{204}Pb$ ratios vary in the studied samples from 10 to 71.5, whereas the radiogenic Pb excess, which accumulated mostly in the L-2 phases, is insignificant. Consequently, the uncertainty in determination of slope ratios for the "isochrons t_1–t_2" is comparable to the determination error for the age t_2. In fact, the slope ratios of these isochrons, which have been calculated on the basis of Pb isotopic data either uncorrected or corrected for the radiogenic Pb accumulation during the last 470 Ma, overlap each other within the limits of uncertainty.

The weighted average age value calculated for dolomites of the Min'yar Formation on the basis of three dates (780 ± 200, 780 ± 110, and 770 ± 200 Ma), which are obtained using different interpretations of Pb–Pb data, corresponds to 780 ± 85 Ma. We consider this value as reliably characterizing the formation age. It is concordant with the formation age limits established on the basis of direct isotopic–geochronological and paleontological data obtained for the Karatau Group, on the one hand, and inferred from the chemostratigraphic telecorrelation of upper horizons of this group.

CONCLUSION

(1) Strict geochemical criteria (Mn/Sr, Fe/Sr, and $\delta^{18}O$) and difference between $^{87}Sr/^{86}Sr$ ratios in the secondary and primary carbonate phases were used to select six rock samples of the Min'yar Formation, which seem to be altered to a minimum extent during the events of burial and meteoric diagenesis; they are considered as the "best" samples from our collection in terms of Rb–Sr systematics.

(2) This preliminary procedure does not guarantee, however, the selection of samples, the Pb–Pb isotopic systems of which are intact, because any epigenetic fluid containing Pb is able to lower the U/Pb ratio and to change the Pb isotope composition in a considerable volume of carbonate rocks, where the Pb concentration is extremely low. The epigenetic Pb–Zn mineralization manifested in the Min'yar Formation is the most distinct indication of such fluid activity.

(3) The isotopic data obtained after separation of epigenetic carbonate phases and sulfide microinclusions by means of leaching in the 0.5 N solution of HBr were used to calculate a series of Pb–Pb dates for the selected dolomite samples. These are the date of 780 ± 200 Ma for the whole rock samples lacking sulfides, the date of 780 ± 110 Ma for primary carbonate phases residual after leaching in 0.5 N HBr, and the date of 470 ± 80 Ma for the acid leachates and residual carbonate fractions of one sample bearing sulfide microinclusions. The last date apparently corresponds to the formation time of Pb-Zn mineralization (t_2). One more date of 770 ± 200 Ma is calculated for the Min'yar dolomites using the "isochron t_1–t_2" model and assuming t_2 = 470 Ma.

(4) The weighted average value for three dates obtained by different approaches is 780 ± 85 Ma. The relatively high uncertainty of this value characterizing age of the Min'yar Formation is conditioned by irregular epigenetic alterations in the studied rocks, the U–Pb and Pb–Pb systems of which have been variably modified. Nevertheless, the indicated date is concordant with the direct isotopic-geochronological and paleontological data available for the Karatau Group, and also with the results of chemostratigraphic correlation based on the C- and Sr-isotope records. The date can be considered as reliably characterizing age of the middle part of the Min'yar Formation.

(5) The sample selection method and analytical procedure used in this work opens a perspective for obtaining the stratigraphically significant Pb–Pb isochron age of rocks, the U–Pb and Pb–Pb systems of which were reset in the course of epigenetic sulfide mineralization.

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The Late Karelian in the Stratigraphic Scale of Russia: Determination of Its Lower Boundary and Regional Units in the Stratotype Area

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Abstract—The U–Pb zircon age is determined for stratigraphically important Upper Karelian structures of the northern Ladoga region. The age of the Sortavala Group volcanics is assumed to be 1.97–1.95 Ga that coincides with age of the Outokumpu and Jormua ophiolites in Finland. The Sortavala and coeval Suisar groups are recommended to be distinguish as the Livvian, the initial regional unit of the upper Karelian, which, along with ophiolites of Finland, marks the opening stage of the Svecofennian belt. The overlying Ladoga Group is a stratotype of the early Ladogian marking the mature evolutionary stage of Svecofennides. The upper boundary of the Ladoga Group is established on the basis of gabbroid intrusions intersecting the group and dated back to 1.88–1.89 Ga. The subsequent late Ladogian interval (up to 1.80 Ga) that characterizes an intense volcanism in the region is missing from the successions.

Key words: **Lower Proterozoic, upper Karelian, Sortavala Group, Ladoga Group, isotope dating.**

INTRODUCTION

The Precambrian stratigraphic scheme of northern Eurasia calles for the two-unit subdivision of the Lower Proterozoic or Karelian, which is assumed to reflect main stages in the Baltic Shield evolution: the early stage of rather stable development and the late stage of the inception and development of the Svecofennian mobile belt. The value of 1900 ± 50 Ma is chosen as the age of the boundary between the early and late Karelian. The boundary postdate the onset of intense volcanism and formation of the early Svecofennian granites ... and separates the Kalevian (the Ladoga Group) from the Vepsian (Semikhatov *et al.*, 1991, p. 8).

In the boundary characterization given above, two phenomena are united: the early Svecofennian volcanism, best manifested and dated in eastern regions of Finland, and the post-Kalevian events terminating the Svecofennian epoch, which separate the Ladoga Group of the northern Ladoga region from the Vepsian. Subsequent investigations showed the necessity to investigate better these phenomena, to verify their position in the evolution of Svecofennides, and to distinguish the standard subdivisions, which determine the boundary between the lower and upper Karelian, as well as the boundaries of regional units. The work is concerned with consideration of these problems on the basis of data on the Sortavala Group metabasites and Ladoga Group metaterrigenous rocks studied in the northern Ladoga region, one of the stratotype regions of the Lower Proterozoic rocks in Russia.

SORTAVALA GROUP

The Sortavala Group exemplifies manifestations of the early Svecofennian volcanism. The geology and chemical composition of the group are extensively studied and described in detail, especially in works by Svetov *et al.* (1990, 1992, and others). The group is exposed in the southern margin of the Archean craton subjected to an intense structural–metamorphic reworking in the Svecofennian time (Fig. 1). It corresponds to a thick, up to 2 km, volcanic pile occurring on the Archean crystalline basement (in places with relicts of the Jatulian, probably upper Jatulian, metasedimentary cover) and underlying metaterrigenous rocks of the Ladoga Group. Metamorphosed basic volcanics transformed into amphibolites and biotite–hornblende crystalline schists dominate in the Sortavala Group; rare thin interlayers are composed of metadacites and metaterrigenous acidic rocks altered to the state of biotite and granite–biotite gneisses. Previously, the Sortavala Group was attributed to the Ludicovian on the basis of the geological correlation, but its isotopic age remained unknown, except for the amphibole K-Ar dates ranging from 1.98 to 1.57 Ga (*Materialy...*, 1982). Numerous metabasite dikes piercing gneisses of the Archean basement are associated with metavolcanics of the Sortavala Group.

North of the Sortavala Group area, metabasites form the independent dike belts, or sills and thin interlayers in rocks of the Jatulian metasedimentary cover. These are better represented and more thoroughly studied in the neighboring territory of Finland (the Koli–Kiihtelysvaara region), where several generations of intraplate continental tholeiites are recognized (Pekkarinen and Lukkarinen, 1991; Vuollo *et al.*, 1992; Nykänen *et al.*, 1994; Kohonen, 1995) in the lowermost Jatulian (2.2 Ga), near the top of the lower Jatulian (the Koljola Formation and others, 2.1 Ga), and at the top of the upper Jatulian (the Ottola Formation, presumably 2.06 Ga old) (Fig. 2). Tholeiites of much later generation (1.97 Ga) are similar in age to the Outokumpu and Jormua ophiolites (1.97–1.95 Ga) thrust over the craton (Koistinen, 1981; Kontinen, 1987; Peltonen *et al.*, 1996). In general, the succession described above can be interpreted as characterizing transition from the stage of the pre-Svecofennian tension of the continental crust to its disruption and formation of Svecofennian ophiolites 1.97–1.95 Ga ago.

Geochronology of rock samples from the Sortavala region were studied in order to refine the position of the Sortavala Group in this succession (Fig. 3). The studied member consists of metabasites ranging in composition from picritic basalt to basaltic andesite and intercalated with rare thin interlayers of metadacites (Table 1, Fig. 4) and metapelites[1] (Table 1). In the chemical composition, the Sortavala Group rocks correspond to intraplate continental tholeiites (Svetov and Sviridenko, 1992), but there are some varieties among them resembling the ocean-floor basalts (OFB) on the basis in the content of rare and trace elements. These rocks were recognized by Ivanikov *et al.* (1997). Judging from lower concentrations of large-ion lithophile elements (LILE) and light rare earth elements (LREE), as well as low Ta/Nb and Th/Yb ratios, the rocks (Table 1, 1956-11, iv) can be considered as primitive in composition as basalts of the Jormua ophiolite association (Peltonen *et al.*, 1996).

Taking into account a low content of syngenetic zircon in basalts, samples for age determinations were taken from thin (several meters) dacite interlayers (Samples 1956-4 and 1645), as well as from the axial part of a thick (about 80 m) metagabbro dike differentiating to quartz diorite (Sample 1672-8). Investigation results for zircons are as follows.

Samples 1956-4 and 1645. Zircon is represented mainly by translucent and muddy prismatic grains of variable length with a sharp or blunt pyramid. Most of grains have rounded edges and apexes, and their faces exhibit traces of dissolution; such grains are often zoned and comprise relic cores of old zircon. Corroded grains represent the first zircon group analyzed (Table 2,

[1] Nearly a third of zircon crystals from metapelites is rounded; their surface shows traces of abrasive scratches and chips. In addition to geochemical peculiarities of metapelites, these features allow a reliable discrimination of metasedimentary rocks from acid volcanics.

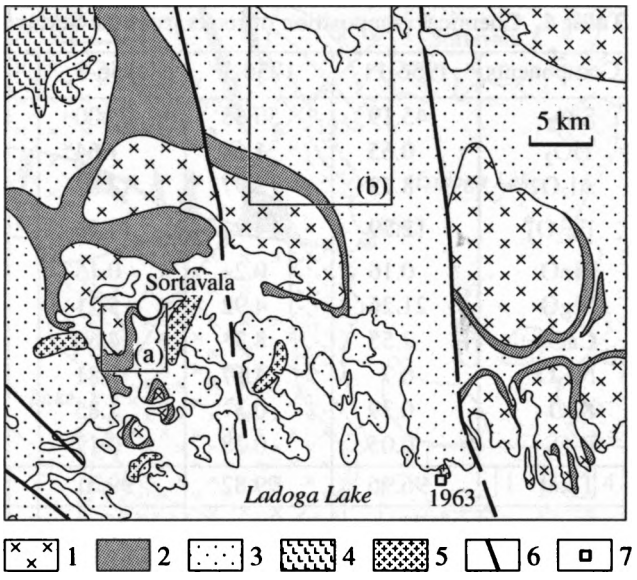

Fig. 1. Schematic geological map of the northern Ladoga region: (1) Archean crystalline basement; (2) Upper Karelian Sortavala Group; (3) Upper Karelian Ladoga Group; (4) gabbro-norites (the Kaalamo Massif); (5) synorogenic granitoids; (6) faults; (7) sampling site in the granitoid Impiniemi Massif (Sample 1963); (a) and (b) areas shown in Figs. 3 and 6.

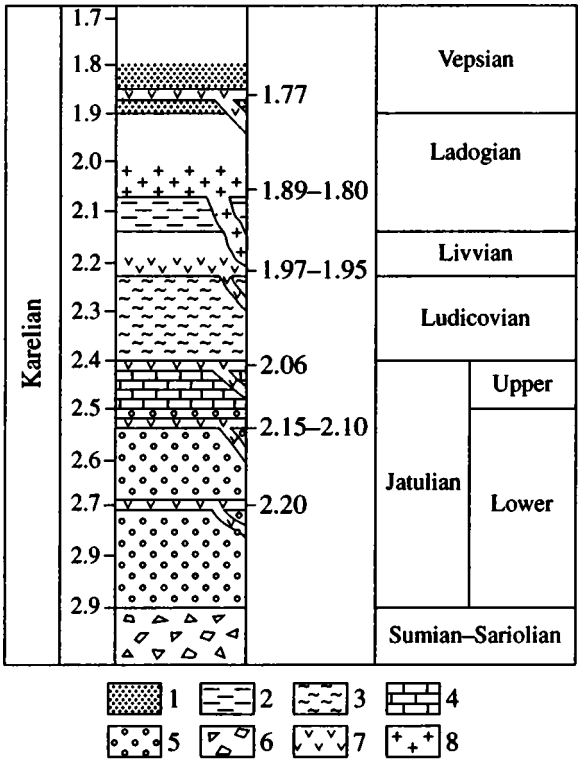

Fig. 2. Lithostratigraphic scheme of the Lower Proterozoic in southwestern Karelia and adjacent regions of Finland (ages, Ma). Characteristic deposits: (1) red beds; (2) rhythmic turbidites; (3) carbonaceous deposits; (4) carbonates; (5) quartzite; (6) coarse-grained sediments; (7) basic volcanics; (8) granitoids.

Table 1. Chemical composition of rocks from the Sortavala Group succession

Components	1956-11	1956-1	1956-3	1645	1956-4	IV	1672-8	MP
SiO_2	45.19	50.63	55.00	63.86	69.88	48.11	46.83	61.46
TiO_2	0.65	1.64	1.64	0.32	0.14	0.83	1.56	1.51
Al_2O_3	8.72	13.97	12.52	18.03	17.21	14.75	7.8	14.13
$Fe_2O_3^*$	12.29	14.9	14.96	3.36	1.83	11.12	16	6.78
MnO	0.16	0.2	0.15	0.04	0.03	–	0.22	0.13
MgO	21.26	4.92	5.21	2.08	0.75	8.01	15.09	2.38
CaO	7.55	8.3	4.69	4.6	3.39	12.74	10.15	3.77
Na_2O	0.7	4.67	1.81	5.58	5.55	1.77	1.38	2.94
K_2O	0.39	0.3	2.85	1.01	0.97	0.1	0.38	2.15
P_2O_5	0.05	0.28	0.17	0.14	0.05	0.07	0.15	0.70
Total	96.96	99.82	99.01	99.03	99.8	97.51	99.56	95.94
Cr	1620	–	142	–	84	240	–	0.02
Ni	1090	–	65	–	48	–	–	–
Co	87	–	49	–	3	47	–	–
Sc	23	–	40	–	2	41	–	–
V	161	–	254	–	6	–	–	0.01
Cu	120	–	76	–	8	–	–	–
Pb	3	–	7	–	18	–	–	–
Zn	77	–	153	–	41	–	–	–
Cs	1.06	–	3.52	–	0.45	–	–	–
Ba	53	–	843	–	845	–	–	0.08
Sr	39	–	191	–	592	125	–	–
Ga	11	–	19	–	16	–	–	–
Li	7.46	–	29.7	–	13.3	–	–	–
Ta	0.14	–	0.61	–	0.26	0.12	–	–
Nb	2.6	–	11.9	–	4.7	–	–	–
Hf	1.11	–	3.52	–	1.91	1.15	–	–
Zr	34	–	142	–	78	32	–	–
Y	14	–	36	–	4	–	–	–
Th	0.3	–	1.86	–	0.75	0.25	–	–
U	0.12	–	0.25	–	0.38	–	–	–
La	3.6	–	19.1	–	8.4	2.36	–	–
Ce	9.05	–	41.9	–	16	6.4	–	–
Pr	1.36	–	5.44	–	1.77	–	–	–
Nd	6.17	–	23.4	–	6.4	4.3	–	–
Sm	1.98	–	5.58	–	0.97	1.65	–	–
Eu	0.72	–	1.6	–	0.62	0.62	–	–
Gd	2.13	–	6.16	–	0.93	–	–	–
Tb	0.37	–	0.96	–	0.1	0.47	–	–
Dy	2.38	–	5.84	–	0.65	–	–	–
Ho	0.47	–	1.18	–	0.12	–	–	–
Er	1.41	–	3.38	–	0.3	–	–	–
Tm	0.19	–	0.51	–	0.05	–	–	–
Yb	1.05	–	3.35	–	0.35	1.73	–	–
Lu	0.18	–	0.47	–	0.05	0.28	–	–

Note: Samples: 1956-11, 1956-1, 3, 4, 1645 represent metavolcanics surrounding the Sortavala Dome; IV, metavolcanics surrounding the Kokkasel Dome (after Ivanikov *et al.*, 1997); 1672-8, metagabbro from dike; MP, metapelite from host rocks of the Sortavala Dome. Analytical methods: X-ray spectral, fluorescent, and instrumental neutron-activation (components in wt %; element concentrations in ppm).

fractions 1, 9–11, 14). Grains with more acute faces and slight corrosion traces belong to the second group (fractions 2–8, 12); they lack zoning, but show old cores.

The $t(^{207}Pb/^{206}Pb)$ ages for all fractions of corroded grains range from 2103.4 to 2452.1 Ma (fractions 1, 8–10, 12). At the same time, the $t(^{207}Pb/^{206}Pb)$ age of zircons belonging to the second group is less than that of zircons from the first group; it varies within a narrower range of 1917.8 to 2051.9 Ma (fractions 2–7, 11). We believe that the variable zircon ages in the second group may be explained by the presence of old cores in the grains.

In the Wetherill diagram, data points of corroded zircons (the first group) from samples 1956/4 and 1645 plot along a single trend approaching the Archean age values. An isochron plotted on the basis of data points characterizing fractions 1, 8, and 10 yields a more exact age of 2917 ± 42 Ma with MSWD value equal to 1 (Fig. 5).

Data points for the second zircon group (fractions 2–7, 12) do not fit the isochron dependence that indicates presence of old radiogenic lead in these zircons. The line connecting two leftmost points (fractions 7, 12) corresponds to the age of 1987 Ma, and this date can be regarded preliminary as the maximum age of the analyzed zircons.

Sample 1672/8. Zircon is represented by small, prismatic, opaque, partially corroded grains. Age values $t(^{207}Pb/^{206}Pb)$ for three fractions vary within a narrow interval: from 1946.5 to 1952.0 Ma (Table 2, fractions 14–16). The corresponding isochron age is 1963 ± 19 Ma, MSWD = 2 (Fig. 5).

According to isochron dating, volcanics of the Sortavala Group appear to be similar in age to the Outokumpu and Jormua ophiolites and younger tholeiitic dikes of the Koli–Kiihtelysvaara region; with due account for the apparent correlation, the age of the Sortavala Group is within the interval of 1.97–1.95 Ga. According to geochemical data, the relevant epoch was distinguished by sharply differentiated geodynamic conditions that resulted in simultaneous formation of OFB and within-plate tholeiites (WPT). Coexistence of both rock types in a single successions of the Sortavala Group seems to be a consequence of their tectonic juxtaposition, when ophiolites were thrust over the continent edge.

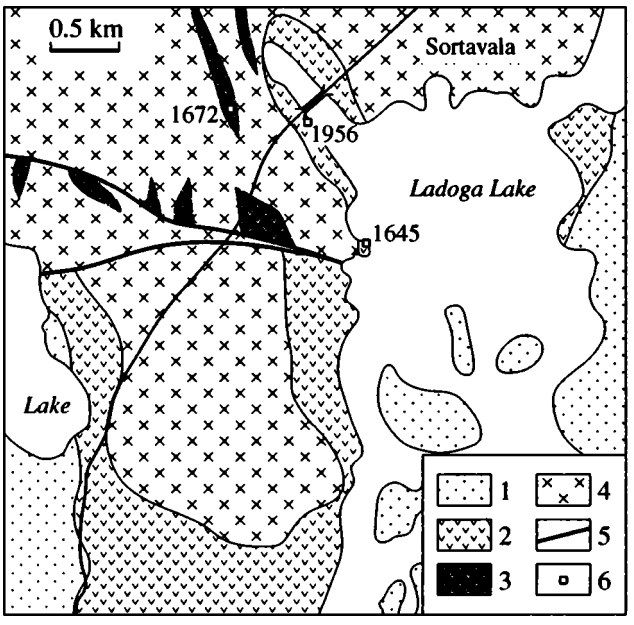

Fig. 3. Geological structure of the southern margin of the Sortavala Dome: (1) Ladoga Group; (2) Sortavala Group; (3) metabasites; (4) granite gneiss of the Archean basement; (5) tectonic dislocations; (6) sites of geochronological samples

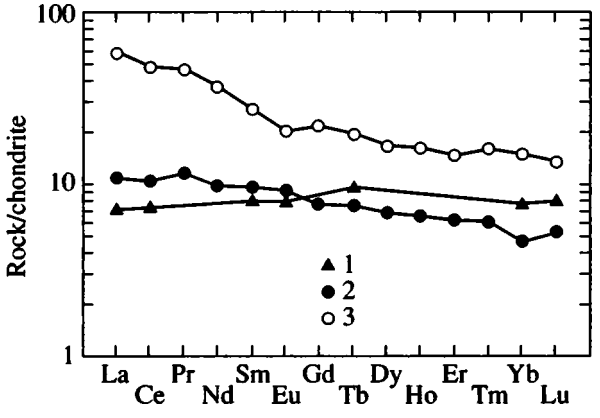

Fig. 4. REE distribution patterns in the Sortavala Group rocks. Samples: (1) 1956-11; (2) 1956-4; (3) 1956-3 (see Table 1)

LADOGA GROUP AND SVECOFENNIAN INTRUSIONS

The Ladoga Group corresponds to a thick (several thousands of meters) sedimentary cover overlaying the Sortavala Group (Predovskii *et al.*, 1967). Intense deformations hinder an unambiguous interpretation of its stratigraphy and relationships with the Sortavala volcanics, though its higher stratigraphic position is evident from the general structure. The group is composed of terrigenous rocks metamorphosed to greenschist and amphibolite facies; turbidites with the rhythmical lamination are very characteristic of the group.

The Ladoga Group is a typical representative of the Kalevian, though its age has been debated over a long time and is disputable until nowadays. Several years ago, most of researchers assumed the relatively old age from 2.0 to 1.9 Ga for the group (Gaal and Gorbatschev, 1987). Later, Huhma *et al.*, (1991) showed that the youngest grains of detrital zircon from the Kalevian rocks are 1.92 Ga old that determined the maximum age limit of sediments, particularly, in the Outokumpu region not far away from the Ladoga Lake region.

Table 2. Pb–U isotopes in zircons

Ordinal no.	Fractions	Weight, mg	U, μg/g	Pb, μg/g	$^{206}Pb^{a}/^{204}Pb$	$^{206}Pb^{a}/^{207}Pb$	$^{206}Pb^{a}/^{208}Pb$	$^{206}Pb^{b}/^{238}U$	$^{207}Pb^{b}/^{235}U$	$t_{(207/206)}$, Ma
Sample 1956/4, metadacite interlayer										
1	>85, l.-pr.	0.70	543	182	3944	7.4869	17.958	0.3270	5.879	2103.4
2	>85, bl.-pyr.	0.09	1365	429	2459	7.9794	20.010	0.3103	5.137	1957.3
3	50–85, l.-pr.	0.61	572	199	3497	7.6778	18.154	0.3405	5.946	2051.9
4	50–85, l.-pr.	0.23	302	95.1	1096	7.6766	15.091	0.3028	4.939	1930.5
5	50–85, l.-pr.	0.36	398	135	873.4	7.1156	12.373	0.3179	5.501	2036.1
6	50–85, pr., ab. tr.	0.28	602	199	703.5	7.0880	9.7162	0.3027	5.102	1989.0
7	50–85, pr., ab. tr.	0.25	385	125	2040	7.9958	15.789	0.3166	5.179	1935.8
8	<50	0.71	467	141	2347	8.0555	20.479	0.2967	4.851	1935.4
9	t. pyr.	0.16	514	218	2222	6.0558	9.3647	0.3868	8.514	2452.1
10	t. pyr.	0.18	203	83.9	1481	6.1716	11.175	0.3817	8.087	2387.0
11	round., transp.	0.18	157	61.9	840.4	6.2878	7.2815	0.3507	6.960	2275.0
12	bl.-pyr.	0.52	279	87.4	1339	7.8571	16.441	0.3039	4.921	1917.8
Sample 1645, metadacite interlayer										
13	small, l.-pr.	0.49	58.0	21.2	299.0	6.0990	5.7733	0.3035	4.999	1948.3
14	>50, pr.	0.18	382	141	2185	7.2222	13.924	0.3531	6.454	2132.4
Sample 1672/8, quartz-diorite from basic dike										
15	<60	0.83	201	88.0	692.5	7.2013	2.8039	0.3318	5.478	1952.0
16	>60	1.18	119	53.1	595.4	7.0598	2.6234	0.3316	5.456	1946.5
17	<60, pr., ab. tr.	0.14	92.8	40.5	527.1	6.9103	2.9783	0.3406	5.618	1950.6
Sample 3055, plagiogranite-porphyry dike										
18	>110, cr. frag.	0.27	60.0	29.9	205.0	4.8739	3.5601	0.3667	7.192	2254.7
19	>110, cr. frag., ab. tr.	0.29	39.0	24.1	486.6	5.0409	2.9816	0.4633	11.07	2589.4
20	>110, pr.	0.80	245	104	2278	6.2036	9.0772	0.3843	8.257	2410.8
21	>110, sh.-pr.	0.49	74.7	236.0	1540	5.742	6.9299	0.4244	9.729	2520.2
22	>75, round.	0.15	219	113	2112	5.7764	4.2603	0.4240	9.788	2531.9
23	75–110, pr.	0.32	107	38.2	1287	6.8514	12.316	0.3380	6.337	2176.6
24	75–110, round., ab. tr.	0.06	355	211	410.7	4.8986	4.3675	0.4683	11.26	2599.6
Sample 260600, gabbro-diorite, Alattu Massif										
25	>75	0.34	165	61.7	1075	7.8285	3.5491	0.3060	4.871	1886.7
26	>75, ab. tr.	0.39	348	139	2705	8.3226	3.7834	0.3350	5.327	1884.9
27	60–75	0.10	337	125	974	7.7485	3.6386	0.3055	4.862	1886.9
28	>75, ab. tr.	1.00	267	111	449.7	6.8361	3.5329	0.3248	5.222	1904.6
Sample 260611, plagiogranite-porphyry dike										
29	>110, pr.	0.50	20.8	10.4	210.3	4.5385	2.8039	0.3513	7.788	2463.6
30	2 grains, ~250	0.10	9.69	9.44	36.96	2.0452	0.8915	0.3028	5.937	2254.0
31	85–110	0.65	15.8	8.52	118.8	3.8293	2.4484	0.3340	7.066	2384.3
Sample 1963, plagiogranite, Impiniemi Massif										
32	<100, pr., ab. tr.	0.76	304	95.1	337.7	6.4978	5.2568	0.2574	4.052	1866.8
33	85–100, t. pyr.	0.48	253	72.9	459.9	7.0387	6.1189	0.2479	3.859	1846.5
34	85–100, bl.-pyr.	0.47	685	154	371.6	6.8918	5.6249	0.1894	2.841	1779.0
35	85–100, pr., ab. tr.	0.65	169	53.6	389.2	6.7924	5.4669	0.2664	4.140	1843.7
36	60–85, cr. ph.	0.77	100	34.8	866.5	7.7115	7.3091	0.3135	4.940	1868.7
37	33–60, cr. ph.	2.40	42.2	14.4	1835	8.2809	8.8030	0.3192	4.997	1857.0
Sample 1091, quartz monzonite, Vuoksa Massif										
38	bulk sample	0.69	773	253	7470	8.9545	5.3879	0.2949	4.474	1799.7
Sample 2199, mafic rocks, Vuoksa Massif										
39	bulk sample	2.14	365	120	6390	8.9179	8.8119	0.3118	4.736	1802.2
Sample 22200, syenite, Vuoksa Massif										
40	bulk sample	1.56	490	162	3726	8.7874	5.8320	0.2974	4.524	1804.6
Sample 10150, quartz monzonite, Vuoksa Massif										
41	bulk sample	1.18	451	161	1416	8.3339	4.4036	0.3047	4.646	1809.3

Note: Grain size, μm; (pr.) prismatic; (l.-pr.) long-prismatic; (sh.-pr.) short-prismatic; (bl.-pyr.) blunt-pyramidal; (t.-pyr.) tapered-pyramidal; (round.) rounded; (transp.) transparent; (ab. tr.) abrasive treatment; (cr. fr.) crystal fragment; (cr. ph.) crystalline phase obtained by Zr dissolution in HF.

Zr decomposition, as well as Pb and U separation is carried out using Krogh methods (Krogh, 1973). The blank run contamination was less than 0.1 ng for Pb and 0.01 ng for U. Pb and U isotope measurements are performed using mass spectrometer MAT-261. Measurement errors for U/Pb isotope ratios do not exceed 0.65% (1.5% for fractions 29–31). Correlation coefficients of measurement errors for Pb/U ratios are from 0.90 to 0.91. All calculations are made using computer programs by Ludwig (1987, 1988).

(a) measured Pb isotope ratios; (b) calculated ratios of radiogenic Pb and U atoms.

Debates continued, and Finnish geologists (Kohonen, 1995) used the term Kalevian as devoid of the stratigraphic status and used to designate the lithologic subdivision; the age of its base in successions near to the Ladoga region was believed to be older than 2.0 Ga.

In the Ladoga region, the Ladoga Group position above the Sortavala volcanics allows us to admit Huhma's interpretation and to consider the age of the group base as younger than 1.92 Ga. Such an interpretation agrees well with the assumed back-arc situation in the basin, where Ladoga turbidites accumulated synchronously with island-arc volcanics known in nearby regions of Finland (Shul'diner *et al.*, 1995, 1996).

The minimum age limit of the Ladoga Group is determined by dates for the cross-cutting Svecofennian intrusions. The youngest intrusions are represented by the Kaalamo gabbro-norite massif and by minor intrusions and dikes of metamorphosed hypabyssal rocks. The latter were originally described by Lobanov (1966) as a regional complex, and Artamonova (1989) united these into the Suistamo Complex in the course of the geological survey. The Kaalamo gabbro-norites are dated back to 1885 ± 5 Ma (Ivanikov *et al.*, 1998). Intrusions of the Suistamo Complex vary in composition from gabbro to tonalite and plagiogranite porphyry. Most of the bodies are represented by dikes, the thickness of which is up to several meters, in places exceeding 50–100 m. They cut across stratified beds of the Sortavala and Ladoga groups, being persistent in composition and exhibiting subvolcanic appearance over the vast territory in zones with a different intensity of folding and metamorphism (Fig. 6). Plagiogranite-porphyry dikes are cut by pegmatitic to aplitic granite veins and subjected to migmatization, according to Lobanov's data. Intrusions are obviously of a synkinematic type, underwent cleavage, and their rocks acquired gneissic structures in zones of a high-grade metamorphic facies.

Gabbroids and plagiogranites reveal similar concentrations and REE distribution patterns (Table 3, Fig. 7a). Different minor intrusions with SiO_2 content varying from 47 to 72% correspond in composition to calc-alkaline K-Na series displaying geochemical characteristics of the 1st type. On the basis of isotopic-geochemical data, the Suistamo Complex can be considered as association, in which granitoids represent the remobilized material of the Archean basement, whereas intermediate rocks correspond to the mixing products of gabbroid and plagiogranite melts.

Gabbroids were sampled in the largest stock located near the Settlement of Alattu (Fig. 6). At the erosion surface, the stock radius is up 0.6 km. Inclination of walls is subvertical. Gabbroids were penetrated by drilling to the depth of 150 m. They enclose steeply dipping plagiogranite-porphyry veins. Relic ophitic and porphyritic textures are rare. Most of rock varieties in the stock are represented by gabbro-amphibolites with directive granoblastic or granoporphyroblastic textures. Dark-colored minerals represented by relic clinopyroxene, hornblende, and subordinate biotite make up 40–80% of rock volume. Zircon and apatite are accessory minerals. Sample 260600 of relatively melanocratic gabbro-diorite was chosen for dating. It yielded colorless, translucent, prismatic zircon crystals with one prism and blunt pyramids. The $t(^{207}Pb/^{206}Pb)$ ages estimated for three zircon fractions vary within a narrow interval from 1884.9 to 1886.9 Ma. The age of isochron plotted across three points is 1884.8 ± 3.2 Ma, MSWD = 0.4 (Fig. 8).

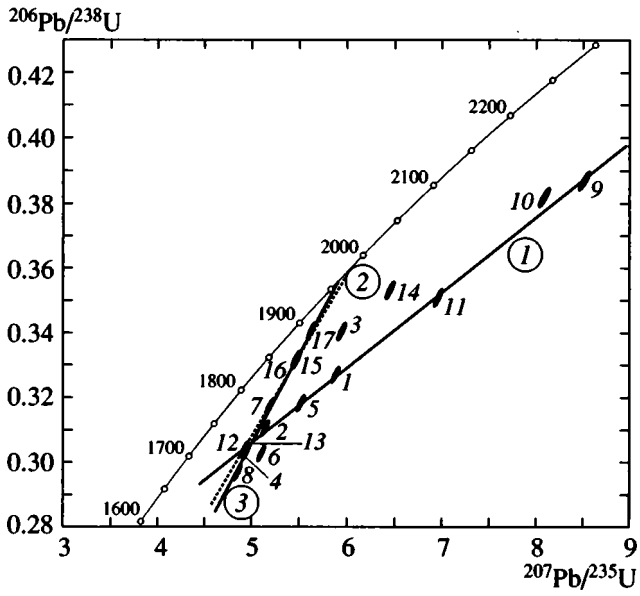

Fig. 5. Wetherill diagram for zircons from the Sortavala Group. Isochrons (encircled numbers): (*1*) the first group of zircons from metadacite interlayers (Samples 1956-4 and 1645); (*2*) the second group of zircons from the same rocks and (*3*) from metabasite dike (Sample 1672-8). Numbers of data points correspond to ordinal numbers of fractions in Table 2.

In contrast to gabbroids, syngenetic Proterozoic zircons have not been found in plagiogranite-porphyry, because the latter probably originated owing to the anatexis of the Archean crustal material. Plagiogranite porphyry from the largest intrusion to the west of the Janisjoki River (Sample 260611) and from its satellite dike (Sample 3055) exposed at the Janisjoki River water edge were sampled for dating procedure (Fig. 6). The intrusion represents a dike-like body 300 m thick, which extends from the north to the south along the distance of 1.5–2 km. The rock texture is porphyritic: tabular plagioclase phenocrysts (up to 1.5 cm long) and rounded quartz phenocrysts (up to 2 cm across) are set in a fine-grained matrix consisting of quartz, plagioclase, and biotite. Cataclasis and gneissic transformations are very intense. In the satellite dike about 10 m thick, plagiogranite porphyry is more fine-grained, with an intense impact of cataclasis. Its porphyritic texture is less distinct. Zircons in the studied samples are

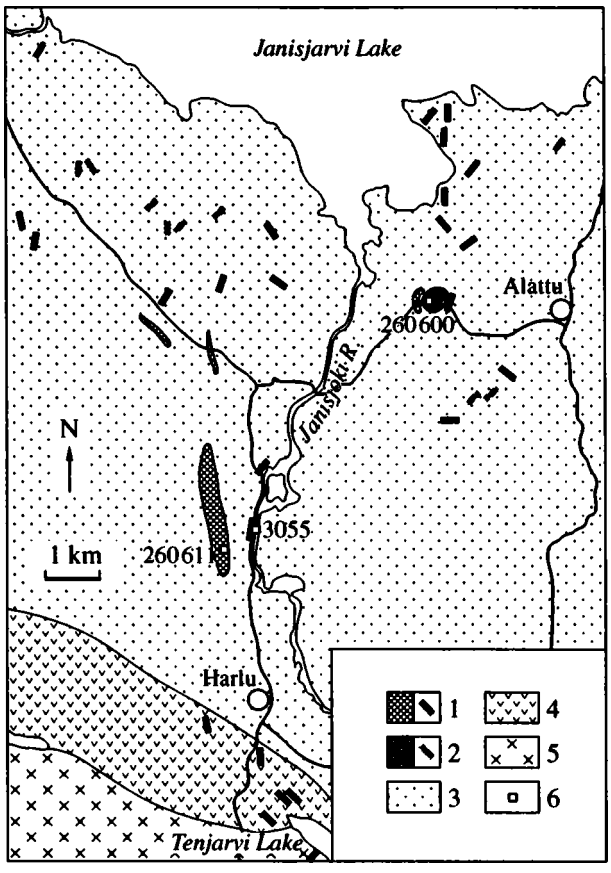

Fig. 6. Distribution of the Suistamo Complex intrusions southward of the Janisjarvi Lake (after K.I. Stepanov): (1, 2) Suistamo Complex of tonalite to plagiogranite porphyries (1) and gabbro-diorite (2); (3) Ladoga Group; (4) Sortavala Group; (5) granite gneiss of the Archean basement; (6) sites of geochronological samples.

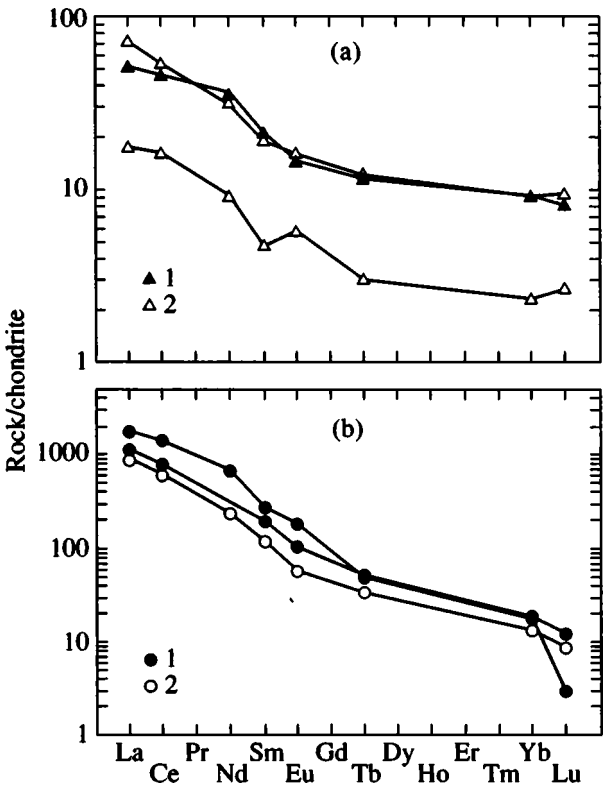

Fig. 7. REE distribution patterns in rocks of intrusive complexes of the Ladoga Lake region. The Suistamo Complex (a): 1, gabbro-diorite; 2, tonalite- and plagiogranite-porphyry. The Vuoksa Massif (b): 1, mafic rocks (samples 2686 and 1463, see Table 3); 2, quartz monzonite.

represented by opaque corroded grains fractured to a different extent and varying from long prismatic to isometric habit. Pyramid tops are sharp to blunt, and crystal edges are rounded. Age values $t(^{207}Pb/^{206}Pb)$ estimated for eight fractions of morphologically different zircons vary from 2176.6 to 2599.6 Ma. In the Wetherill diagram, data points for both samples plot along two isochrons having different upper intersections with concordia (Fig. 8). The isochron age of zircons from Sample 3055 is calculated to be 2705 ± 60 Ma (MSWD = 20), and that for zircons from Sample 260611 is 2850 ± 95 Ma (MSWD = 0.89).

Hence, the upper age limit of the Ladoga Group is about 1.89 Ga, as it follows from dates obtained for the Kaalamo and Suistamo complexes of gabbroid intrusions. The formation time of the latter coincides with that of the basin closure during the accretion of an island-arc system (Nironen, 1997) and marks the beginning of a long period, when the Svecofennian plutonic complexes were formed.

The subsequent synorogenic granitoids, the origin of which is related to the main stage of folding and metamorphism, are established to be 1.87–1.86 Ga old in the western Ladoga region (Kozakov *et al.*, 1997). In the northern Ladoga region, the Impiniemi massif occupying the synonymous peninsula at the Ladoga Lake coast and some neighboring islands is the best known among synorogenic plutons. The observed contacts of the massif with enclosing schists of the Ladoga Group are distinctly intrusive. At the same time, granitoids and enclosing rocks are concurrently metamorphosed into gneissoid medium-grained rocks. Tonalites and plagiogranites dominant in the massif are mainly composed of plagioclase, quartz, biotite, and occasional microcline (Lobach-Zhuchenko *et al.*, 1974). Zircon, sphene, and apatite are accessory minerals. The Impiniemi plagiogranites exhibit geochemical characteristics of the 1st type (Table 3). Sample 1963 (Table 2) of the leucocratic granite from the Ruotsinsari Island located west of Impiniemi Cape was dated (Fig. 1). Zircon of the sample is represented by prismatic well-cut crystals with traces of slight dissolution during secondary process. By means of selective dissolution and aeroabrasion, grains with slightly disturbed U–Pb isotope relationships was separated. On the basis of the isochron method, zircons (Table 2, fractions 33–37)

Table 3. Chemical composition of intrusive rocks

Components	260600	MB-51	3055	MB-50	260611	MB-11/4	1963	2686	2199	1463	22200	1091	10150
SiO_2	50.40	50.80	61.43	62.20	64.30	67.10	72.70	39.30	45.17	48.54	52.93	58.17	60.23
TiO_2	1.20	1.30	0.81	0.41	0.31	0.23	0.21	2.17	2.96	2.16	1.45	1.16	1.14
Al_2O_3	15.30	14.50	15.98	16.70	15.40	16.40	14.34	12.20	15.80	16.26	18.30	16.95	16.17
FeOt	11.48	11.39	5.45	5.27	4.09	2.63	2.20	9.39	11.06	10.64	8.08	6.10	5.33
MnO	0.17	0.19	0.06	0.14	0.07	0.04	0.03	0.18	0.17	0.14	0.12	0.08	0.07
MgO	5.80	5.10	4.60	1.30	3.50	1.70	0.75	10.92	5.96	4.22	3.23	2.75	2.69
CaO	7.70	8.20	4.69	5.40	5.20	4.40	1.53	14.05	9.93	7.14	4.88	4.19	3.98
Na_2O	2.40	2.50	3.01	4.00	3.80	4.60	4.33	1.60	3.85	3.40	4.59	4.14	3.55
K_2O	2.30	1.90	2.04	2.10	1.30	0.95	3.02	3.13	1.78	3.57	5.06	4.35	5.15
P_2O_5	0.35	0.39	0.18	0.29	0.16	0.11	0.05	4.13	1.49	1.56	0.85	0.62	0.68
Total	98.30	99.37	99.78	100.21	98.90	99.76	99.67	98.99	98.83	98.39	100.07	99.65	99.23
H_2O								0.42	0.06	0.22	0.04	0.08	0.30
F								0.25		0.17	0.11		0.16
Ba	714	700	571	680	810	623	1346	8260		3400	5280		3475
Zr	81	140	152	142	65	78	99	342		496	560		534
Rb	68	55	149	69	36	49	80	94		73	111		154
Sr	641	523	525	909	793	761	315	4087		3656	3759		1912
Co	28	31.0		11.0	14.0	7.2	3.2	28.0		18.0			11.0
Y	18		21	20	7	<15	18	40		78	37		36
Nb	<10		15	<10	<10	<10	16	20		32	21		32
Cs	2.0	3.5		5.0	5.4	1.5	1.4						
Cr	61.0	38.0		16.0		55.0	8.2	55.0		4.3			24.0
Sc	31.0	2.8		8.6	11.0	4.9	3.3	19.0		16.1			10.0
Hf	1.6	3.7		4.2	1.9	2.4	3.1	6.1		4.3			11.4
Th	1.5	2.6		3.5	0.5	0.6	4.3	40		26	36		36
U	0.4	0.7		0.4	0.5	0.7	1.4	6.4		1.4			
Ta	0.2	0.3		0.5	0.2	0.2	1.3	0.5					2.1
La	12.0	17.0		24.0	8.7	5.8	24.0	566.3		357.0	550.0		282.4
Ce	24.0	39.0		46.0	14.0	14.0	40.0	1199.0		631.0	690.0		507.0
Nd	16.0	22.0		20.0	6.6	5.7	11.0	407.0					146.0
Sm	3.3	4.3		3.9	1.4	1.0	1.9	51.8		39.1			23.9
Eu	1.1	1.1		1.2	0.5	0.4	0.6	13.6		7.8			4.4
Tb	0.5	0.6		0.6	0.2	0.2	0.2	2.3		2.4			1.7
Yb	1.7	2.0		2.0	0.7	0.5	0.8	3.8		3.9			2.9
Lu	0.3	0.3		0.3	0.1	0.1	0.1	0.1		0.4			0.3

Note: Samples: gabbro-diorites (260600, MB-511), tonalite porphyries (3055, MB-50), and plagiogranite-porphyries (260611, MB-11/4) of the Suistamo Complex; granite of the Impiniemi Massif (1963); apatite-bearing calc-alkaline lamprophyre (2686), melanodiorite and monzodiorite rich in apatite and magnetite (2199, 1463), syenite (22200), and quartz-monzonite (1091, 10150) of the Vuoksa Massif. Analytical methods: X-ray-spectral, fluorescent, and instrumental neutron-activation.

were dated back to 1874 ± 24 Ma (MSWD = 14), if fraction 32 is excluded from calculations (Fig. 9).

Postorogenic intrusions represent an important reference complex, the terminal one in the Svecofennian magmatic cycle. These small magmatic bodies of a complicate composition form a peculiar sublatitudinal belt, which extend from the western Ladoga region to the Aland Islands of Finland. The age of postorogenic intrusions in the Ladoga region was estimated using the Vuoksa Massif as an example. The massif is a large ring

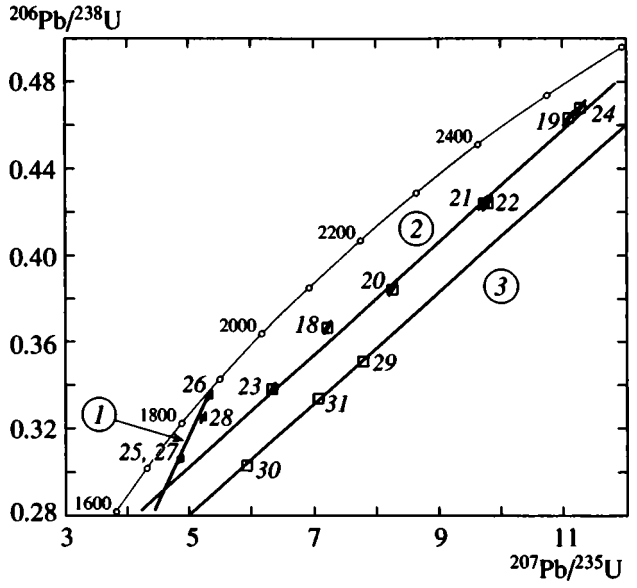

Fig. 8. Wetherill diagram for zircons from the Suistamo Complex. Isochrons (encircled numbers): (*1*) zircons from gabbrodiorite (Sample 260600); (*2, 3*) zircons from plagiogranite porphyry (samples 3955 and 260611). Numbers of data points correspond to ordinal numbers of fractions in Table 2.

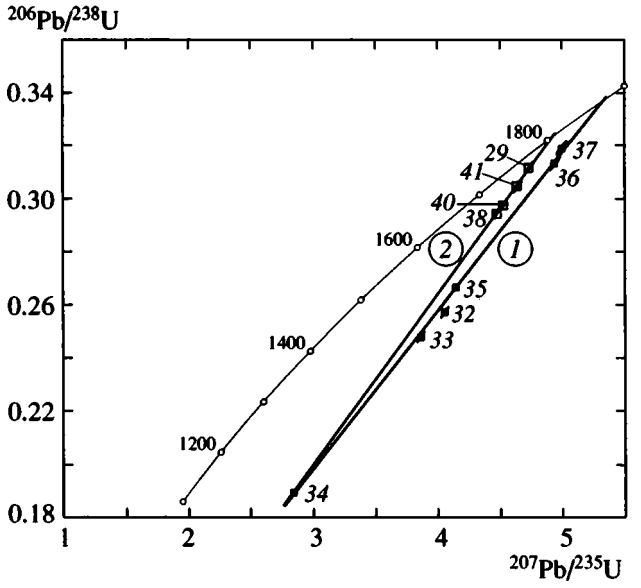

Fig. 9. Wetherill diagram for zircons from intrusive massifs. Isochrons (encircled numbers): (*1*) for synorogenic Impiniemi (Sample 1963) and (*2*) postorogenic Vuoksa (samples 1091, 2199, 22200, and 10150) massifs. Numbers of data points correspond to ordinal numbers of fractions in Table 2.

intrusion composed of rocks ranging in composition from apatite-bearing calc-alkaline lamprophyres, ultramafics, and subalkaline gabbroids to quartz monzonites and granites (Ivanikov *et al.*, 1995; Konopel'ko, 1997; Konopel'ko *et al.*, 1998). Geochemical features of postorogenic rocks, such as high concentrations of K_2O, P_2O_5, REE, and volatile elements versus relatively low contents of Ta, Nb, Ti, Cr, Ni (Table 3), and low $^{87}Sr/^{86}Sr$ initial values correlate them with shoshonite-latite series in interpretation of Tauson (1982). The REE distribution patterns are highly fractionated but lacking negative Eu anomalies (Fig. 7b).

The age of the massif was determined on the basis of several zircon samples. Zircons were separated from rocks of contrast composition (Tables 2, 3): from melanocratic diorite (Sample 2199) and mesocratic syenite (Sample 22200) of the 1st intrusive phase, and from quartz monzonites of the 2nd intrusive phase (samples 1091 and 10150). Ages $t(^{207}Pb/^{206}Pb)$ estimated for these samples vary within a narrow interval from 1799.7 to 1809.3 Ma (Table 2, fractions 38–41). In the Wetherill diagram (Fig. 9), data points can be approximated by a line, the upper intersection of which with concordia corresponds to the age value of 1807 ± 44 Ma (MSWD = 10). The considerable error presumably reflects a certain difference between the formation times of the 1st and 2nd intrusive phases.

The origin of postorogenic complexes is related to the uplift movement and subsequent onset of the platform stage. A quiet epoch after their formation resulted in accumulation of the Vepsian platform red beds preserved at the recent denudation level in the Onega region. Accordingly, the estimated age value of 1.8 Ga may correspond to the termination of the pre-Vepsian stratigraphic hiatus. The Vepsian quartzite–sandstone enclose the concordant dolerite bodies 1.77 Ga old (Bibikova *et al.*, 1990), and the base of the Vepsian thus corresponds in age to a date within the interval of 1.8–1.77 Ga.

The study on Nd isotope composition in magmatic rocks of the northern Ladoga region (Table 4) showed that

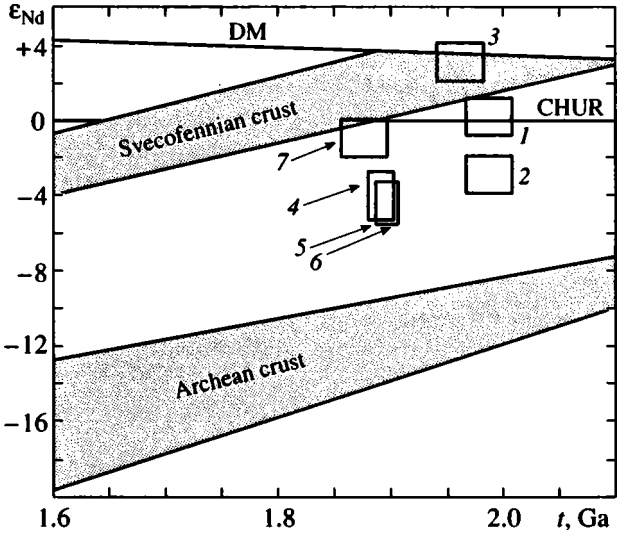

Fig. 10. ε_{Nd}–*t* diagram for rocks of the northern Ladoga region. Samples: (*1*) 1956-1; (*2*) 1956-4; (*3*) 1672-5; (*4*) 260600; (*5*) 260611; (*6*) 3055; (*7*) 1963 (see Table 4).

Table 4. Sm–Nd data for bulk-rock samples from the northern Ladoga region

Sample no.	Rocks	t, Ma	Sm ppm	Nd ppm	$\frac{^{147}Sm}{^{144}Nd}$	$\frac{^{143}Nd}{^{144}Nd}$	2s	$\varepsilon_{Nd}(0)$	$\varepsilon_{Nd}(t)$	t_{DM}, Ma	t_{DM}, Ma
1956-1	Metabasalt	1990	5.15	21.30	0.1464	0.511985	10	–12.7	0.1	2630	2450
1956-4	Metadacite	1990	0.90	5.33	0.1025	0.511261	13	–26.9	–2.8	2580	2690
1672-5	Gabbro	1963	3.05	9.59	0.1930	0.512738	16	2.0	2.9	3030	2200
1963	Plagiogranite	1874	1.59	9.30	0.1035	0.511437	13	–23.4	–1.0	2360	2450
260600	Gabbro-diorite	1885	3.67	16.60	0.1339	0.511656	10	–19.2	–4.0	2840	2700
260611	Plagiogranite porphyry	1885	1.34	6.69	0.1212	0.511473	14	–22.7	–4.5	2750	2740
3055	Plagiogranite porphyry	1885	0.87	4.64	0.1141	0.511384	12	–24.5	–4.5	2690	2740

Note: Measured $^{143}Nd/^{144}Nd$ ratios are normalized to $^{143}Nd/^{144}Nd = 0.511860$ of the La Jolla standard. Calculations of t_{DM} are after Wasserburg *et al.* (1981), and t_{DM_2} after Liew and Hofmann (1988).

the rocks reveal peculiarities typical of many rocks in the Baltic Shield, which are different in age and have mantle origin, as the ε_{Nd} parameter changes from values typical of the depleted mantle to the negative ones (Fig. 10). It is believed that such a decrease in ε_{Nd} is best explained by the contamination of magmas (the source of which is the depleted mantle) with the material of an old continental crust rich in LREE (Andersen and Sundvoll, 1995). The model of this kind was suggested, for instance, in Finland (Patchett and Kouvo, 1986) where Svecofennian rocks appear to be contaminated by the Archean crustal material. When applied to the studied rocks, the model well agrees with presence of the Archean zircon xenocrysts distinguished in the rocks.

DISCUSSION

The data presented shows that the evolution of the northeastern continental margin of the Svecofennian belt was controlled by three principal events: (a) by the early tholeiitic volcanism at the stage of the continental crust rifting and ophiolite formation 1.97–1.95 Ga ago; (b) by subsequent accumulation of the Ladoga Group deposits during the time interval of 1.92–1.89 Ga ago, which corresponds to the stage of island-arc volcanism in internal parts of the belt (at the territory of Finland); and (c) by formation of plutonic complexes within the interval of 1.89–1.80 Ga ago.

If the beginning of the Svecofennian epoch corresponds to the lower boundary of the upper Karelian, then the boundary age should be established on the basis of the ophiolite formation period 1.97–1.95 Ga ago. As evident from our data, this event corresponds to the accumulation time of the Sortavala Group tholeiites in the northern Ladoga region that represents a marginal zone of the Karelian craton and to the formation of the volcanic–plutonic association of picritic basalts of the Suisar Group (Formation) in a further internal part of the craton, namely, in the Onega region, where it originated 1974 ± 27 Ma ago according to the Sm–Nd systematics (Pukhtel' *et al.*, 1995). Thus, we turn back to the idea of Bibikova *et al.* (1989), who suggested to define the boundary between the lower and upper Karelian by the commencement of the Svecofennian events and to date it back to 1.97 Ga. There are two advantages of this idea as compared to the level of 1.90 Ga adopted the stratigraphic scale of 1990: first, the older date is consistent with natural periodicity in the geological history and marks such a significant event as the opening of the Svecofennian Belt; second, it is based upon dating for the stratigraphic unit (Sortavala Group), whereas the date of 1.90 Ga splits the Svecofennian evolution at the most significant point and corresponds to the age of plutonic processes that contradicts principles of chronostratigraphy. In the scale of 1990, both of the advantages were sacrificed in favor of correlation ability between the East European and Siberian subdivisions. At present, however, this argument turns out to be insignificant, because events, unknown earlier and revealed now in the Precambrian of Siberia, coincide in age with the commencement of Svecofennides evolution (Gavrikova *et al.*, 1991; Kotov *et al.*, 1993; and others). Data on such events are scarce and inadequately known, but they signify an important turning point in the Proterozoic evolution of Siberia. In general, the level of 1.95 Ga is regarded at present as one of the most important datums in the Early Precambrian history, which is traceable in all continents (Glebovitskii and Shemyakin, 1996).

In the current regional scheme, the Sortavala and Suisar groups are included into the Ludicovian, and, from this standpoint, the latter combines two units different in age and lithology: the metasedimentary carbonaceous shales of the Onega Formation or their analogs and the overlying metavolcanics of the Suisar Group and, as is now established, of the coeval Sortavala Group. Materials discussed above indicate a necessity of excluding the mentioned volcanogenic series from the Ludicovian like it was proposed earlier by V.A. Sokolov and supported by A.P. Svetov, who con-

sidered the Suisar volcanics as rocks of the independent Livvian "period" (*Materialy...*, 1982; Svetov and Sviridenko, 1992). Following this proposition, we may consider the Livvian as the regional subdivision corresponding to the Sortavala Group and beginning the upper Karelian succession. The exact geochronometric constraints of the Livvian are not established, but we may assume that its lower boundary is approximately 1.97 Ga old, whereas the upper boundary age is inferable from the onlap of the Ladoga Group, the base of which is not older than 1.92 Ga (with due account for data by Huhma *et al.*, 1991). It should be pointed out that a bit older island-arc basalts and andesites (1.93–1.91 Ga) were revealed in adjacent regions of Finland (Ekdahl, 1993; Nironen, 1997; and others). Bearing in mind the geodynamic considerations, genetic relationships, and almost the same age of island-arc volcanics and back-arc turbidites, we may assume that these and those should be united into a single stratigraphic unit with the lower boundary dated at 1.93 Ga ago, but this is still a problem to be verified in future.

According to our data, the top of the Ladoga Group representing the Kalevian is not younger than 1.888–1.885 Ma (dates for intrusive rocks of the Kaalamo and Suistamo complexes). In the regional scheme, the Svecofennian plutonic complexes are considered as post-Kalevian, characterizing the next Vepsian unit. Such a decision appears to be inadequate, first, because it contravenes to the rule of establishing the stratigraphic boundary by the base of the overlying unit, and second, because it displaces the typical orogenic Svecofennian complexes and the post-Svecofennian red-bed series of the Vepsian age. To avoid these misunderstandings, some authors suggest that an independent Svecofennian unit should be distinguished between the Kalevian and Vepsian. This is not the appropriate decision, since the stratified rock succession of that age is unknown in the region, and the unit name turns out to be coinciding with the name of a longer Svecofennian tectonic epoch. Taking into consideration this and the ambiguous interpretation of the term Kalevian in recent works of Finnish researchers, we should not use the term in the Russian stratigraphic scale. The interval between the top of the Livvian and the base of the Vepsian could be regarded as a regional unit named Ladogian (this has been once proposed in the literature). With such a solution of the problem, the pre-Vepsian stratigraphic hiatus and concurrent plutonic complexes can be attributed to the late Ladogian.

CONCLUSION

New results of isotopic dating of some Karelian rocks in the northern Ladoga region show that the position of boundary separating the lower and upper Karelian in the general stratigraphic scale of the Precambrian in Russia should be revised with the proper modification of the regional subdivision scheme of the upper Karelian. From this viewpoint, we suggest the following:

(1) to define the boundary between the lower and upper Karelian by the basic event of the epoch—the commencement of the Svecofennian Belt evolution, using the age information for the Outokumpu and Jormua ophiolites in neighboring regions of Finland and for the Sortavala Group volcanics in the Ladoga region;

(2) to unite volcanogenic basaltoid pile of the Sortavala Group in the Ladoga region with the Suisar Group of the Onega region and to consider this combination as rocks of regional Livvian unit, the lower boundary of which is 1.97 Ga old;

(3) to use the name Ladogian for the overlying subdivision of the regional scale within the time-range from 1.92 Ga (age of detrital zircons from the concurrent subdivisions in Finland) to 1.80–1.77 Ga ago (basal age of the Vepsian). The stratotype of its lower part is the Ladoga Group, the revised upper age limit of which corresponds to about 1.885 Ga. The younger interval is not represented in the stratigraphic succession as it corresponds to orogenic and postorogenic plutonic complexes of Svecofennides.

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Middle Devonian and Frasnian Seas of Siberia

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Abstract—The paleolandscape schemes for the Eifelian, Givetian and Frasnian seas of Siberia are considered for the first time in the present work. On the basis of sediment lithology, composition of benthic communities, their habitat environments, and sedimentation conditions, the territory of Siberia is divided into fourteen Eifelian paleolandscapes comprising twelve particular regions, twenty one Givetian paleolandscapes divisible into four regions, and fifteen Frasnian paleolandscapes also subdivided into four regions. The paleolandscapes are characterized by data on their geographic position, size, sedimentation trends, relief and substratum of sea bottom, hydrodynamics, water salinity and temperature, tectonic regime, and volcanic activity. The comprehensive analysis of benthic populations provided grounds to conclude that the distinguished marine basins of Siberia were freely connected with nearby and distant seas, and this was favorable for migration of various organisms. It is also established that the widespread biohermal and reefal buildups extended like a wide belt from the Altai and Salair seas across the Nyurol'ka–Var'egan paleolandscape to the north of the West Siberian sea. In general, the considered seas were shallow, located in the past subtropical zone. Western and easternmost areas of the West Siberian sea were involved into intense tectonic movements accompanied by submarine eruptions of basic to intermediate lavas in the west and of intermediate to acidic lavas in the southeast.

Key words: **paleolandscape, Eifelian, Givetian, Frasnian, age, paleobasin, land mass, corals, bryozoans, algae, benthos, sediments, volcanics, reefs, bioherms.**

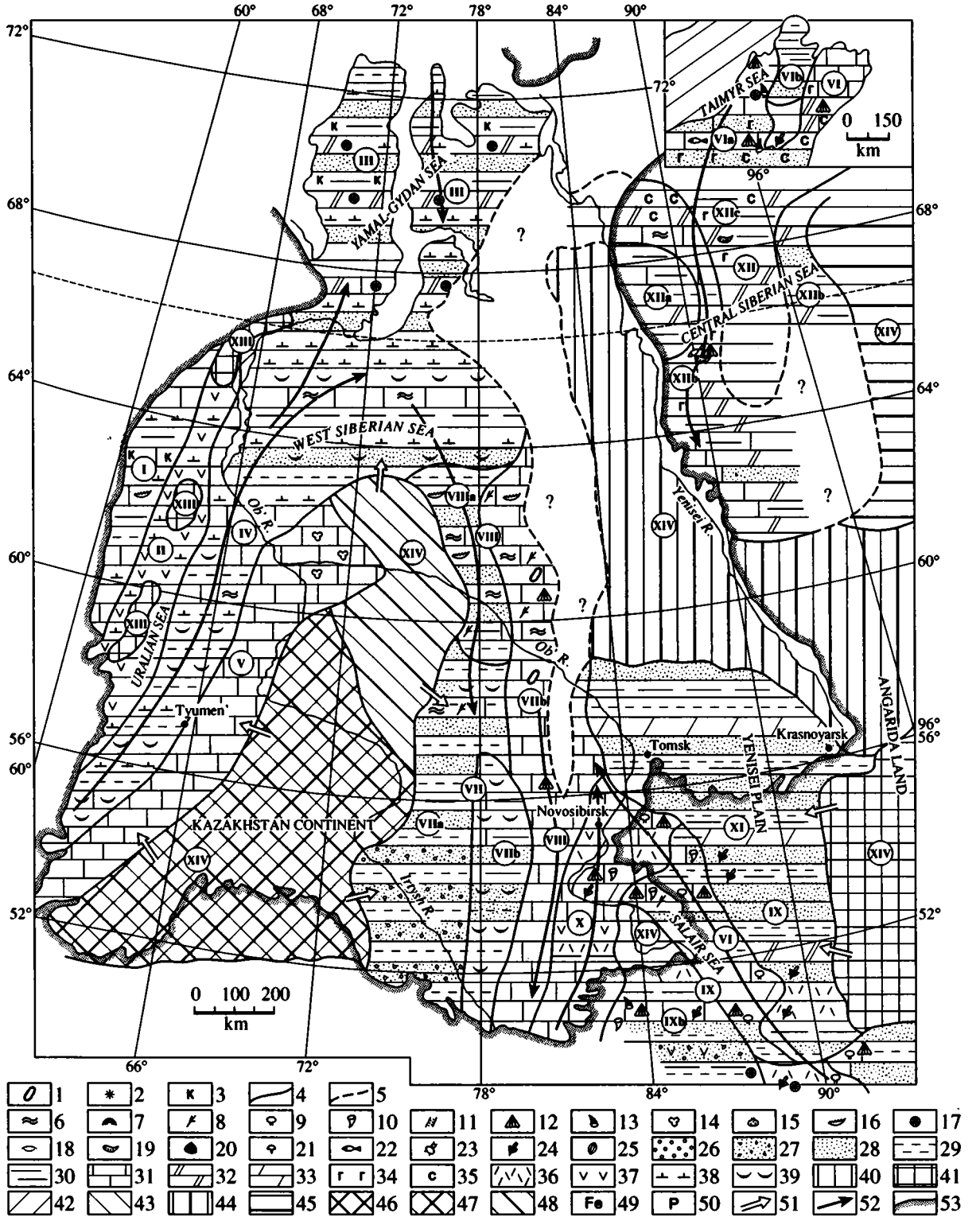
This work is a continuation of the previous one (Dubatolov and Krasnov, 1998), where we considered paleolandscapes of the Early Devonian seas in Siberia. Accordingly, we omit here our explanation of terms "paleolandscape," "landscape," and "paleolandscape region," as it has been given before. Reconstructing the paleolandscapes of Siberian seas, we used the available data of paleontology, stratigraphy, and lithology, the key scientific branches applicable in paleobasin investigations (Sokolov, 1987).

The paleolandscape differentiation of Siberian seas during the Eifelian Age was as complex as during the Emsian. At the time in question, marine sedimentation settings still occupied a vast territory providing favorable habitat environments for various organisms. Volcanic eruptions of basic to intermediate lavas were active as before in the western part of the West Siberian Plain. Only areas surrounding the Kazakhstan continent were devoid of volcanism manifestations, and the continent proper juts out into the West Siberian sea extending northward to the Ob' River middle courses and further. The Yenisei accumulative pain bounded the sea in the southeast. There were 14 paleolandscapes divisible into 12 particular regions during the Eifelian Age (Fig. 1).

The Cis-Urals paleolandscape occupied the westernmost part of the West Siberian sea and accumulated calcareous to siliceous ooze, silt, clay, sand, and basic volcanics. Tectonic movements were moderately active. Sedimentation environments and active hydrodynamics apparently were unfavorable for distribution of organisms. Only rare massive tabulate corals *Favosites goldfussi* Orb. and brachiopods *Bornharditina* sp. and *Zdimir* ex gr. *pseudobaschciricus* (Tschern.) inhabited this part of the sea. However, benthic population in the Uralian sea located eastward was abundant and very diverse. It is likely that an intense influx of sedimentary material was responsible for the water haziness in the Uralian paleolandscape, where the temperature regime was suitable for thermophilic organisms. The water salinity was normal, as is evident from presence of stenohaline tabulate corals.

The East Uralian paleolandscape represented a zone of very active submarine eruptions of intermediate to basic lavas. This zone of the northeastern strike was presumably bounded by faults and active in tectonic aspect.

The Yamal–Gydan paleolandscape situated northward of the West Siberian sea represented a marine basin that accumulated calcareous ooze, siliceous clay, sand, and sporadic intercalations of basic lavas. It is likely that the tectonic activity here was not intense. The terrigenous influx was irregular, slowing sometimes and giving place to accumulation of siliceous-clayey ooze at the bottom. Seawater was clearer in these periods, but salinity remained normal. Small-


60°
68°
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78°
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56°
52°
96°
66°
TAIMYR SEA
YAMAL-GYDAN SEA
WEST SIBERIAN SEA
URALIAN SEA
CENTRAL SIBERIAN SEA
KARA SEA
KAZAKHSTAN CONTINENT
ANGARIDA LAND
YENISEI PLAIN
Ob' R.
Yenisei R.
Irtysh R.
Tyumen'
Tomsk
Novosibirsk
Krasnoyarsk
0 150
km
0 100 200
km

sized crinoids *Salairocrinus* sp. and *Cycloceocrinus* sp. populated the basin.

In the east, the East Uralian paleolandscape joined the Shaim–Yarudei paleolandscape, where the accumulation of silt, clay, sand, basic tuff, and calcareous ooze was concurrent to eruptions of basic lavas. This was typical of a larger, especially northern part of the basin. The poor population of the basin mostly consisted of bryozoans and algae suggesting a shallow depth range (up to 40–50 m). The water salinity was likely close to normal, and temperature did not exceed 20°C. Hydrodynamics was fairy intense, associated with a moderate tectonic activity, as it can be inferred from a rather frequent intercalations of basic lavas in the sedimentary sequence.

Thus, the East Uralian and Shaim-Yarudei paleolandscapes, two major in the West Siberian sea, represented a vast zone of submarine volcanism. The volcanism was most active in the former, where it could negatively influence the development and distribution of marine organisms.

The Ishim–Khanty-Mansiisk paleobasin was situated between the Shaim–Yarudei paleolandscape and the Kazakhstan land mass. Geological data characterizing this paleobasin are scarce, though we can state that this part of the West Siberian sea accumulated calcareous ooze. The northwestern part of the Kazakhstan continent adjacent to the Ishim–Khanty-Mansiisk paleolandscape represented a low denudation plain that provided an insignificant terrigenous influx. Additional data suggest that the northern paleolandscape area was inhabited by foraminifers *Irregularina karlensis* Viss., *Tubeporina gloriosa* Pron., *Cribrosphaeroides* sp., *Uralinella* aff. *bicamerata* E. Byk., *Archaesphaera* ex gr. *minima* Sul., and others.

The Tevriz paleolandscape adjacent to the southeastern margin of the Kazakhstan continent corresponded to an accumulative plain displaying a characteristic differentiation of sediments. It comprised the Dem'yanka and Tui–Barabinsk regions, the first of which accumulated silty, sandy, and pebbly deposits within a low denudation plain. In contrast, calcareous ooze, silt, and sand, the tuffaceous varieties included, represented sediments characteristic of the second, Tui–Barabinsk paleolandscape region. These sediments were deposited in shallow marine settings under conditions of a moderate tectonic activity. The Kazakhstan land mass served as a provenance of a terrigenous material.

The Nyurol'ka–Var'egan paleolandscape occupied the junction area of the Vakh and Ob' rivers approximately in the middle courses of the latter. The Var'egan region of this paleolandscape was situated near the Kazakhstan continent, whereas the narrow Nyurol'ka gulf run deep inland of southern continental area (Fig. 1).

The Var'egan paleolandscape region was distinct owing to composition of sedimentary sequence consisting of limestone, clay, siltstone, and sandstone beds. Predominance of terrigenous rocks here shows that the local sedimentation was under control of erosion in the eastern area of the Kazakhstan continent. Characteristic of the Nyurol'ka paleolandscape region was accumulation of calcareous ooze, tabulate coral pavements, and development of reefal and biohermal buildups. The benthic population was impoverished. Typical of the Nyurol'ka–Var'egan paleolandscape are the following fauna groups: stromatoporoids *Amphipora blokhini* Yavor., *A. pinguis* Yavor., *A.* ex gr. *angusta* Lec., *A. pervesiculata* Lec., and less diverse and massive *Gerronostroma* cf. *uralense* Yavor. (Makarenko, 1991), branching tabulate corals, algae, tentaculitids, and conodonts. Relatively abundant were *Cladopora microcellulata* Dubat. and *Alveolitella karmakensis* (Tchern.) among corals; *Nowakia lautentalensis* Alberti, *N. sulcata* Roemer, *N. karpinskii* Ljasch., *Viriatellina hollardi* Lardeux, and *Styliolina decurtata* Bouček among tentaculitids; and pelagic chitinozoans *Hoegishaera glabra* Staplin. The basin was relatively shallow (as deep as 40 m). Its water clarity favored development of tabulate corals, and salinity was normal, as one can judge from presence of stenohaline stromatoporoid and coral species.

Fig. 1. Scheme of Eifelian paleolandscapes in Siberia: (1) reefal and biohermal massifs; (2) red beds; (3) siliceous deposits; (4) established boundaries between paleolandscapes and their regions; (5) presumable boundaries of this kind; (6) algae; (7) stromatoporoids; (8) ramose tabulate corals; (9) massive tabulate corals; (10) rugose corals; (11) bryozoans; (12) brachiopods; (13) tentaculitids and dacryoconarids; (14) foraminifers; (15) radiolarians; (16) conodonts; (17) crinoids; (18) bivalves; (19) ostracods; (20) phyllopods; (21) trilobites; (22) fishes and Agnatha; (23) eurypterids; (24) terrestrial plants; (25) miospores; (26) shingle; (27) gravel; (28) sand; (29) siltstone; (30) clay; (31) calcareous ooze; (32) dolomitic ooze; (33) marly ooze; (34) gypsum or anhydrite; (35) halite; (36) acidic lavas; (37) lavas of intermediate composition; (38) lavas of basic composition; (39) tuffs of variable composition; (40) erosion land areas of calcareous–clayey sandstones and crystalline rocks; (41) erosion land areas of siliceous slates; (42) erosion land areas of calcareous slates; (43) erosion land areas of aphyric basalts; (44) erosion land areas of greenschists; (45) erosion land areas of clayey–calcareous greenstones and other crystalline rocks; (46) presumable erosion land areas of sandstones and calcareous or siliceous slates; (47) presumable erosion land areas of sandstones, conglomerates, gravelstones, siltstones, and shales; (48) presumable erosion land areas of limestones; (49) iron ores; (50) phosphorites; (51) transport directions of terrigenous material; (52) directions of sea currents; (53) present-day contour of the West Siberian plain.

Paleolandscapes: (I) Cis-Urals; (II) East Uralian; (III) Yamal–Gydan; (IV) Shaim–Yarudei; (V) Ishim–Khanty-Mansiisk; (VI) Taimyr; (VII) Tevriz; (VIII) Nyurol'ka–Var'egan; (IX) Altai–Salair; (X) Kolyvan'–Tomsk; (XI) Yenisei; (XII) Central Siberian; (XIII) island arcs; (XIV) Kazakhstan, Angarida, and Barnaul land masses. Paleolandscape regions: (VIa) West Taimyr; (VIb) Central Taimyr; (VIc) East Taimyr; (VIIa) Dem'yanka; (VIIb) Tui–Barabinsk; (VIIIa) Var'egan; (VIIIb) Nyurol'ka; (IXa) Salair; (IXb) Altai; (XIIa) Igarka; (XIIb) Tunguska; (XIIc) Central Siberian.

The Nyurol'ka paleolandscape region predominantly accumulated carbonate sediments, sometimes together with clayey mud, especially in the northern part. In general, calcareous ooze filled up depressions of slightly undulating bottom relief. The following fauna groups populated these depressions: foraminifers *Auroria sphaerica* Sub., *Bykovaella aperturata* (Pron.), *Irregularina obscura* Reitl., *I. karlensis* Viss., *Bisphaera elegans* Viss., *Parathurammina aperturata* Pron. and others (Zadorozhnyi, 1987); colonies of stromatoporoids *Stromatopora* sp., *Actinostroma mamontovi* Yavor., *A. salairica* Yavor., *A. pervesiculata* Lec., *A. clathratum* Nich., *A. densatum* Lec., *Stachyodes constulata* Lec., *Amphipora ramose* (Phill.) and others (Makarenko, 1991); abundant tabulate corals *Favosites goldfussi* Orb., *Pachyfavosites polymorphus* (Goldf.), *Emmonsia aspera* Yanet, *Thamnopora beliakovi* Dubat., *Striatopora schandiensis* Dubat., *Alveolitella karmakensis* (Tchern.), *Coenites flexibilis* Sok., *C. bulvankerae* Dubat., *Tyrganolites altaicus* Tcherep., and *Syringopora yavorskyi* Tchern.; rugose corals *Tabulophyllum schluteri* (Peetz); brachiopods *Desquamatia pesterevskensis* Ržon., *Undispirifer undiferus* (Roem), and others; fairly uniform bryozoans *Semicoscinium* cf. *altshedaicum* Nekh., *Klausena* ex gr. *immortalis* Tr., and others. Ooze accumulating areas were habitats of tentaculitids *Nowakia sulcata* Roem., *N. karpinskii* Ljasch., *Viriatellina hollardi* Land., *Styliolina decurtata* Bouček, *S. minuta* Bouček (Meretskaya, 1993), and also of ostracods *Bairdiocypris gerassimovi* (Rozhd.) and *Microcheilinella mendymensis* Rozhd. (Savina, 1993). Chitinozoans *Hoegisphaera glabra* Staplin, *Cyathochitina* sp., and others (determinations of N.M. Zaslavskaya) inhabited the pelagic zone. The bottom was covered with the stromatoporoid–coral pavement and with reefal and biohermal buildups. The high diversity of benthic population in the Nyurol'ka paleolandscape region and presence of cosmopolitan species suggest that the West Siberian sea had free connections not only with neighboring, but also with remote seas, the species interchange between which was intense. The water mass was warm (not cooler than 20°C) and sufficiently transparent. The basin depth ranged from 10 to 60 m, never exceeding 100 m (Chuvashov, 1963; Poyarkov, 1977; Dubatolov, 1982). Sedimentation developed under tranquil tectonic conditions, because the fine horizontal lamination of sediments is intact. In general, the Nyurol'ka–Var'egan paleolandscape was very favorable for life activity and dispersal of benthic organisms, whose population in the basin was extremely diverse.

The Altai–Salair paleolandscape included two respective sea basins connected with the southeastern area of the West Siberian sea. Principal types of deposits in this paleolandscape were calcareous ooze, silt, clay, subordinate marl, and acidic volcanic rocks (about 10% of the sequence). The Salair and Altai basins reveal different sedimentation patterns and contrasting communities of benthic organisms. For instance, the extremely diverse benthic communities populated substratum of clayey–silty–calcareous ooze in the Salair region. At the same time, the background accumulation of clayey–calcareous ooze in Altai region was accompanied by deposition of mudstones, siltstones, sandstones, and pebblestones. This situation and concurrent eruptions of acidic lavas in southern areas of the Altai sea were responsible for a considerable diversity of benthic faunas.

The Salair region was inhabited by foraminifers *Cribrosphaeroides simplex* (Reitl.), *Bisphaera elegans* Viss., *Salpingothurammina tuberculata* (Lip.), *Suleimanovella paulis* (E. Byk.), *Bykovaella aperturata* (Pron.), *Kalijanella novita* (Pron.), and others (Zadorozhnyi, 1987), and also by stromatoporoids *Actinostroma mamontovi* Yavor., *Stromatopora dimorphica* V. Khalf., *Stromatoporella undata* Yavor., and others. (*Biostratigrafiya paleozoya...*, 1960). Tabulate corals were represented by 49 species of the widespread Eifelian genera *Favosites, Pachyfavosites, Thamnopora, Cladopora, Alveolites, Caliapora,* and *Placocoenites*, but chaetetiods and heliolitids were less abundant. They coexisted with representatives of genera *Striatopora, Coenites*, and *Tyrganolites*. The low abundance rate was typical of genera *Squameofavosites, Emmonsia, Pleurodictyum, Fomitchevia, Crassialveolites, Egosiella, Natalophyllum, Syringopora, Heliolites*, and *Paraheliolites*. Dominant species of tabulate corals were *Pachyfavosites polymorphus* (Goldf.), *Striatopora schandiensis* Dubat., *Thamnopora alta* (Tchern.), *Th. parva* Yanet, *Alveolites insignis* Tchern., *Caliapora uralica* Yanet, *Coenites bulvankerae* Dubat., *Paraheliolites rarus* (V. Khalf.), *Chaetetes ninae* Tchern., *Ch. rotundus* Lec., and *Chaetetipora possima* Dubat. (Dubatolov, 1959, 1963, 1972). *Columnaria rhenana* Frech, *Neospongophyllum variabile* Wdkd., *N. crassiforme* Wdkd., *Calceola sandalina* Lam., and some other species represented the diverse rugose corals (Cherepnina, 1967; Elkin *et al.*, 1982). Other diverse groups were brachiopods *Gypidula globosa* (Schnur), *Uncinulus* cf. *parallelepipedus* (Bronn.), *Atrypa schandiensis* Ržon., *Emanuella* cf. *subumbona* (Hall), *Lazutkinia mamontovensis* (Laz.) (Rzhonsnitskaya, 1952, 1968) and ostracods *Akidellina* sp., *Aparchitellina monocornis* (L. Egorova), *Knoxiella* aff. *selebratis* (Pol.), *Sulcocavellina* aff. *altchedatensis* Pol., and others (Elkin *et al.*, 1987). Less abundant in the region were trilobites *Schizoproetus salairicus* (N. Tchern.), *Gerastos peetzi* (N. Tchern.), *Proetus lazutkini* N. Tchern., *Astycoryphe senckenbergiana* R. et E. Richter (Elkin *et al.*, 1982), bivalve species *Glossites concentricus* (Goldf.), *Spenotus* cf. pelagica Goldf., *Paleonella* cf. *plana* Hall, and *Nucula lineata* Phill. (Kulikova, 1965), and gastropods representing genera *Bucanopsis, Murchisonia*, and *Strobeus*. At the same time, crinoids *Triacrinus lutulentus* J. Dubat., *Stylocrinus tabulatus tabulatus* (Goldf.), *S. tabulatus depressus* (Müller), *Gasterocoma mite* J. Dubat., *Myrtillocrinus orbiculatus* J. Dubat., *Aporetocrinus ligatus* (Quenst.), *Salairocrinus cingulatus*

(Goldf.), *Cupressocrinites gracilis* Goldf., *C. scaber* Schultze, *Hexacrinites humilicarinatus* Yelt., and *H. tuberosus* Yelt. represent a considerable part of the fossil community (Dubatolova, 1971, 1977). Remains of fishes *Arthrodira* (Ivanova *et al.*, 1964) and *Acanthodes dublinensis* Stauff. along with Crossopterigii and Dipnoi fragments are scarce here (Elkin *et al.*, 1986, 1987). Ammonoids *Pinacites jugleri* (Roem.), *Foordites* sp., and *Agoniatites* sp. inhabited pelagic settings (Bogoslovskii, 1982).

The above species composition suggests that the arid climate and typical marine regime were dominant in the west of the Altai–Sayan province, for instance, in the Salair sea, where the water temperature was not lower than 18–20°C, and the depth range was from 20 to 50 m, as one can judge from buried algae, exotic inclusions in the coral polypary, and thick-walled brachiopods that occur in fossil communities (Dubatolov, 1982).

The species diversity is indicative of favorable habitat environments, e.g., of well-illuminated water mass with normal salinity, moderately intense hydrodynamics, and sufficient nutrient reserves. It was important that the neighboring, gently inclined land surface supplied the Salair sea with organic matter and remains of plants that grew in marshy areas of the southwestern Angarida. The plants were represented by *Psilophyton salairicum* Anan. et S. Step., *Psilophytites rectissimus* Höeg, *Uralia camdjalensis* Petros., *Barrandeinopsis beliakovi* Krysht., *Aneurophyton germanicum* Kr. et Weyl., *Oriensphyton jakubovi* S. Step., *Herbula marinae* S. Step., *Protolepidendron scharianum* K. (Mamontovo Horizon) and *Blasaria sibirica* (Krisht.) Zal., *Aphyllopteris thenuis* Petr. (Akarachkino Horizon; Petrosyan, 1973,1983).

To the southwest of the Salair sea, there was the Altai paleolandscape region, where calcareous to clayey and marly ooze, silt, and gravel had a higher content of terrigenous components. Predominant sediments in the Mongolian Altai area were silt, sand, gravel, and shingle. Sedimentation here was associated with eruptions of acidic lavas, and all lithological characteristics point to the general shoaling of the Altai sea, where shoals and islands were distinguishable. Communities of benthic organisms in the sea were not very diverse because of water pollution by abundant terrigenous material. This concerns first of all an area corresponding to the present-day Altai Mountains. The basin that was located at the site of the present-day Altai ore district was characterized by sedimentation of pure calcareous ooze under conditions of calm hydrodynamics. The transparency of water mass was favorable for development of diverse benthic faunas, in particular, for the appearance of coral pavement at the bottom.

Benthos of the Altai sea (area of the Altai Mountains) was rich in massive and ramose tabulate corals *Favosites goldfussi* Orb., *Pachyfavosites macrotrematus* Dubat., *P. markovskyi* Sok., *P. polymorphus* (Goldf.), *Thamnopora alta* (Tchern.), *T. pulchra* (Tchern.), *Alveolites insignis* Tchern., *Paraheliolites rarus* (V. Hallf.), *Chaetetes magnus* Lee, *Ch. tenuis* Frech; rugose corals *Xystriphyflum sibiricum* (Bulv.), *X. uralicum* (Soshk.), *Taimyrophyllum carinatum* (Bulv.), *Lynelasma halliafdrme* (Soshk.), *L. orientalis* (Soshk.), *Neocolumnaria devonica* Bulv., *Dendrostella vulgaris* (Soshk.), *D. rhenana* (Frech) and others (Cherepnina, 1967, 1969; Gumerova, 1996) along with bryozoans *Lioclema salairiense* Moroz., *Pseudobatostomella salairiensis* Moroz., *Semicoscinium ubensis* Nekh., *S. striatum* Nekh. and others were abundant as well (Volkova, 1974). Brachiopods were less diverse, represented by *Lazutkinia mamontovensis* (Laz.), *Camarotoechia lazutkini* Ržon., *Schuchertella altaica* Khalf. (Gratsianova and Kul'kov, 1960; Stepanov *et al.*, 1972), and trilobites *Neoprobolium altaicum* (Khalf.) and *Scutellum elegans* (Peetz) were scarce (Elkin, 1973). Crinoids *Cupressocrinites gracilis* Goldf., *Hexacrinites humilicarinatus* Yelt., *H. tuberosus* Yelt., and *Mediocrinus microgrumosus* J. Dubat. represent a considerable part of the fossil community (Dubatolova, 1971, 1973, 1977).

Thus, it is evident that the benthic population in the Altai paleolandscape region was less diverse than that of the Salair region, and consequently, the habitat environments in the former were less favorable. Nevertheless, the westernmost areas of the Altai sea (the present-day southern Altai and ore district) were quite suitable for marine organisms. In addition to widespread stromatoporoids, a high diversity (44 species) was typical here of tabulate corals representing genera *Favosites, Pachyfavosites, Emmonsia, Pleurodictyum, Roemeripora, Thamnopora, Striatopora, Cladopora, Rhachopora, Alveolites, Alveolitella, Crassialveolites, Caliapora, Coenites, Placocoenites, Tyrganolites, Natallophyllum, Syringopora, Thecostegites, Aulopora, Mastopora, Heliolites,* and *Paraheliolites.* Genera *Favosites, Pachyfavosites, Thamnopora, Striatopora, Coenites, Placocoenites*, and *Tyrganolites* were dominant. Abundant rugose corals (21 species) represented genera *Syringaxon, Barrandeophyllum, Oligophyllum, Ridderia, Pseudopetraia, Tabulophyllum, Calceola, Nardophyllum, Zonophyllum, Pseudozonophyllum, Digonophyllum*, and *Phillipsastraea* (Spasskii, 1960, with additions of Dubatolov). Bryozoans (22 species) from genera *Fenestella, Lioclema, Fistulipora, Discritella, Crustopora, Hemitrypa, Semicoscinium, Polypora*, and *Rhombopora* (determinations of K.N. Volkova) were relatively abundant. Brachiopods represented 26 species of 22 genera (Bublichenko, 1974), but if we take into consideration additional data obtained recently (Zheltonogova *et al.*, 1994), they, like the corals, were even more diverse. According to Zinevich (1989), crinoid community in the paleolandscape under consideration included 36 species of 21 genera.

The western area of the Altai sea was a habitat of 18 trilobite (Maksimova, 1960) and 2 tentaculitid species (data of V.L. Klishevich). Conodont taxa known from

this area are *Icriodus difficilis* Ziegler, Kappler et Johnson, *I. obliquemarginatus* Bisch. et Zieg., and *Polygnathus linguiformis bultincki* Weddige; algae are represented by *Lancicula alta* Masl., and *Ikella vermiculata* Shuys. (determinations of V.A. Luchinina). Pelagic settings were habitats of chitinozoans *Lagenochitina brevicervicata* Coil. et Schwalb., *Angochitina boyetae* Cramer, *Eisenackchitina aranea* (Urban), and *E. bohemica* (Eisenack) (determinations of N.M. Zaslavskaya), and also of ammonoids *Anarcestes (Latananarcestes) noeggerati* (Buch), *Werneroceras ruppachense* (Kayser), *W. socolicum* Bogosl., and *W. altaicum* Bogosl. (Bogoslovskii, 1958).

Thus, the habitat environments of organisms in the Salair–Altai sea were somewhat different. They were most favorable in the Salair region that was connected with the Nyurol'ka–Var'egan region of the West Siberian sea and located far away from provenances of terrigenous material. Surrounding land areas represented a low alluvial plain, whose contribution to supply of sedimentary material was insignificant. The calm hydrodynamics was favorable for plankton development that provided nutrients for benthic organisms. As a result, different groups of abundant benthic community populated basins situated at the sites of the present-day Salair Ridge and southern Altai Mountains, the ore district included, but they did not penetrate into eastern areas of the Altai–Sayan region. The fauna dispersal was favored by the appropriate water temperature (not less than 18–20°C), and also by the perfect aeration of water mass in the shallow basin, as it follows from algae remains buried in abundance.

The peculiar Kolyvan'–Tomsk paleolandscape was situated between two others, each characterized by different tectonic activity: the Nyurol'ka–Var'egan paleolandscape located westward was passive in tectonic aspect, whereas the volcanic eruptions of acidic lavas were typical of the easterly Altai sea (ore-bearing and central districts of the Altai Mountains). The Kolyvan'–Tomsk paleolandscape represented a zone of moderate tectonic activity, and the background accumulation of clayey to calcareous ooze in this zone developed simultaneously with eruptions of intermediate to acidic lavas. This inhibited evolution of marine organisms so that the paleolandscape was almost devoid of them. Only its northern and northeastern parts were populated by rare brachiopods in association with tabulate and rugose corals. Coral species identified here are *Pachyfavosites* sp., *Thamnopora* sp., *Crassialveolites crassus* (Lec.), *Placocoenites* sp., *Heliolites vulgaris* Tchern., *Chaetetes* sp., *Stringophyllum* sp., and *Grypophyllum* sp.

The Taimyr paleolandscape is distinctly divisible into the western, central, and eastern regions (Fig. 1). The West Taimyr region accumulated calcareous to clayey ooze, sands, and evaporites. Its southeastern and northeastern areas were populated predominantly by stenohaline brachiopods, corals, crinoids, trilobites, and tentaculitids, the dwellers of marine water with normal salinity. Species abundant among brachiopods were *Undispirifer undiferus* (Roem.), *Lazutkinia* sp., *Gypidula biplicata* Schnur, *Clorindina eifelensis* Nal., and *Dentatrypa kolymensis* (Nal.).

The Central Taimyr region was distinct owing to accumulation of silty–clayey deposits in the subtidal zone of the sea, as is evident from coastal plant remains buried in sediments. Among those plants, there were *Protocephalopteris* sp., *Psilophiton* sp., *Taeniocrada decheniana* (Goepp.) Kr. et Weyl., and some other taxa.

Deposition of sediments in the East Taimyr region progressed under conditions of the arid climate, when the temperature of quickly evaporating water could be much higher than 20°C, and the basin depth was not greater than 2–5 m. What should be noted as well is the high water salinity favorable for dispersal of typical marine organisms.

The same conditions were characteristic of the Central Siberian paleolandscape, where the predominant dolomitic deposits turned out to be intercalated with abundant gypsum and halite interbeds. Only a central part of this basin (near the town of Igarka) was devoid of evaporites and accumulated clayey to silty deposits and dolomitic to calcareous ooze. This situation apparently was responsible for the almost complete absence of typical marine dwellers, except for scarce fishes, ostracods, lingulids, and charophytes adapted to the abnormal water salinity in the southwestern area of the Central Siberian sea. Environmental changes toward the normal salinity favored appearance of immigrants, for instance, of tabulate corals *Pachyfavosites polymorphus* (Goldf.), *Thamnopora proba* Dubat., *Gracilopora* sp., *Chaetetes* sp., and also of foraminifers *Parathurammina* cf. *aperturata* Pron., *P.* cf. *tuberculata* Lip., *P.* ex gr. *cushmani* Reitl., *Bisphaera* cf. *irregularis* Bir., and others (Matukhin *et al.*, 1995).

Considering in general the Siberian paleolandscapes of the Eifelian Age, one can easily see that their arrangement was quite regular. First of all, they surrounded the Kazakhstan continent that jutted out far into the West Siberian sea. Western and northern areas of this sea were peculiar because of eruptions of intermediate and basic lavas. The central area of the sea, especially to the east of the Kazakhstan continent, was passive with respect to volcanic activity. Finally, the vast Yenisei alluvial-diluvial plain arose in southeastern regions of the West Siberia and accumulated the terrigenous material derived, most likely, from the Angarida. The Eifelian Age was also the period of initial formation of biohermal massifs (Kolyvan'-Tomsk and Taimyr paleolandscapes) and also of reefal buildups in the Salair (Pesterevo and Belyi Kamen' reefs) and Altai basins (Kholozovo and Losikha reefs in the Altai ore-bearing district). The reefal buildups are also known in southeastern areas of the West Siberian sea (Malyi Ich reef), and all the buildups seemingly formed a narrow reefal belt extending northward from the Salair and Altai seas across the Nyurol'ka region into the Taimyr

sea (Krasnov and Stepanov, 1979; Krasnov *et al.*, 1980; Stepanov, 1970; Stepanov *et al.*, 1984, 1990).

The Givetian paleogeographic zoning was different from that of the Eifelian Age. The main factor responsible for this was the corresponding transgression that flooded vast areas in Siberia and reached for the first time the Taimyr, Central Siberian, and Salair–Minusinsk regions. The Givetian paleolandscape differentiation was more pronounced than that of the Eifelian time, and, in this case, we distinguished 21 paleolandscapes, two of which were subdivided into four regions (Fig. 2). At the time in question, a stable tectonic activity was characteristic of western areas of the West Siberian sea, where volcanic eruptions of variable composition continued since the Early Devonian time to form thick lava sequences.

Characteristic of the Cis-Urals paleolandscapes was accumulation of rudaceous terrigenous deposits and calcareous ooze. Areas with the terrigenous substratum were basically populated by brachiopods and foraminifers. In the Nizhnii Tagil region, the borehole sections are rare, and data available are insufficient for depicting the complete distribution pattern of benthic organisms. In the larger Uralian sea that was located westward, the Givetian benthic population was very diverse and included, in addition to brachiopods, the abundant corals, ostracods, and crinoids. There were also widespread dozens of foraminifer species, for instance, *Bykovaella bykovae* (Pojark.), *Auroria ferganensis* Pojark., *Cribrosphaeroides robusta* M.-Maclay, and *Bisphaera malevkensis* Bir. included (Zadorozhnyi, 1987). They associated with the following taxonomic groups: stromatoporoid species *Actinostroma bifurcata* Nich., *A. stellulata* Lec., *A. septate* Lec., *Columnostroma fenestrata* (Nich.), *Trupetostroma porosa* Lec., *Stachyodes verticulate* (McCoy), *Amphipora spica* Yavor., and others, more than ten in total; chaetetids *Pachytheca irregulare* Yanet and others; more than 20 species of tabulate corals, the widespread *Thamnopora nicholsoni* (Frech), *T. bublichenkoi* Dubat., *T. cervicornis* (Blainv.), *Gracilopora vermicularis* (McCoy), *Dendropora dubrovensis* Dubat., *Alveolites taenioformis* Schlut., *Alveolitella fecunda* (Salee sensu Lec.), *Crassialveolites obtortus* (Lec.), *Caliapora battersbyi* (M.-Edw. et H.), *C. taltiensis* Yanet, *Coenites tenella* Gurich, *Placocoenites gradates* (Lec.), *Scoliopora denticulate* (M.-Edw. et H.), *S. conferta* Erm., and *Tyrganolites eugeni* Tchern. included. The sea under consideration was also populated by 16 rugose corals representing genera *Thamnophyllum, Pachyphyllum, Pseudomicroplasma, Stringophyllum, Bethanyphyllum, Neostringophyllum, Heliophyllum, Hexagonaria, Peneckiella, Spongophyllum, Grypophyllum, Alaiophyllum*, and *Tabulophyllum*, and also by more than 20 brachiopod species representing 16 genera (*Kishechnopolostnye...*, 1972). Twelve ostracod species represented genera *Amphissites, Geisina, Cavellina, Bairdiocypris, Microcheilinella, Bairdia*, and *Samarella.* Species of crinoids (Dubatolova, 1971) and conodont (Snigireva and Shuiskii, 1977) were also diverse. However, an impoverished benthic population was typical of the western periphery of the West Siberian sea that was a part of the adjacent Cis-Urals paleolandscape. Brachiopods from this area were represented by *Stringocephalus burtini* Defr. and *S. manjensis* Khod. et M. Breiv., and foraminifers by *Parathurammina bykovae* Pojark. and *P.* sp. It is likely that scarcity of brachiopods and foraminifers, as well as absence of other faunas, reflect unfavorable conditions for benthos development within the paleolandscape that quickly accumulated terrigenous deposits, mostly tuff, gravel, and shingle. The basin depth was here not greater than few dozens of meters.

Volcanic activity in the East Uralian paleolandscape was inherited from the Eifelian volcanism. Lava flows of intermediate and basic composition accumulated here against the background sedimentation of clayey ooze. Data on organic life in this zone are unknown. As we believe, the benthos penetration into the paleobasin was prevented by active tectonics and volcanism that caused the intense hydrodynamic perturbations.

According to scarce data on sediments in the Yamal-Gydan paleolandscape located northward of the West Siberian sea, they were deposited under extremely quiet tectonic environments of the platform-type. The paleolandscape accumulated siliceous clay, calcareous to dolomitic ooze, sand, and basic volcanic rocks. Data on organic life here are scarce despite the obviously favorable habitat environments. Fossil groups suggesting these environments are small-sized brachiopods of the family Spirifiridae, ramose stromatoporoids, and tabulate corals representing families Auloporidae, Aulocystidae, Alveolitidae, and Coenitidae, among which species *Alveolitella grata* Isaev was most abundant.

The Shaim paleolandscape extended as a narrow zone eastward of the East Uralian paleolandscape. In our opinion, it was a submeridional tectonic zone bounded by deep faults. The basin floor was covered here with calcareous ooze populated by bryozoans. Tectonic conditions were similar to those in the neighboring East Uralian and Kustanai–Tyumen paleolandscapes. The water temperature in the basin was about 18–20°C. The graben-like structure of the zone remote from provenances of terrigenous material and the thermal regime in the basin apparently were suitable for accumulation of calcareous ooze.

Further to the east, there was the Kustanai–Tyumen paleolandscape that formerly represented a marginal lowland of the Kazakhstan continent and turned into the alluvial-diluvial plain in the Givetian time. The rock succession includes here sandstone, gravelstone, and pebblestone beds intercalated with interbeds of tuff and lavas variable in composition. Thus, in tectonic aspect, this was an active zone of a vast alluvial denudation plain. The northern part of the zone extended into the West Siberian sea, whereas the southern part was outside the sea within the territory of present-day Kazakhstan.

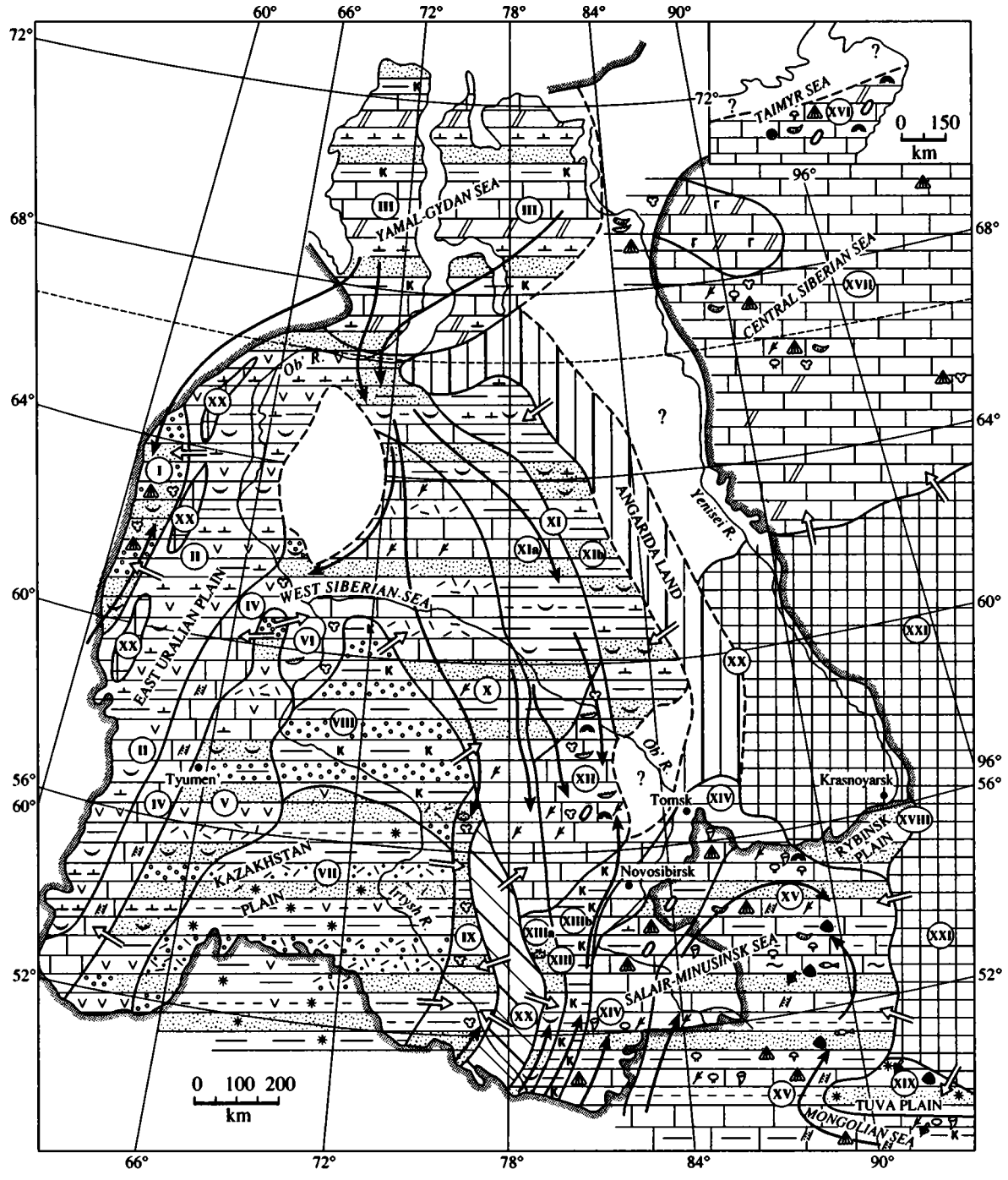

Fig. 2. Scheme of Givetian paleolandscapes in Siberia (symbols as in Fig. 1). Paleolandscapes: (I) Cis-Urals; (II) East Uralian; (III) Yamal–Gydan; (IV) Shaim; (V) Kustanai–Tyumen; (VI) Khanty-Mansiisk; (VII) Kazakhstan; (VIII) Dem'yanka; (IX) Tevriz; (X) Nyurol'ka–Var'egan; (XI) Vakh–Nadym; (XII) Sil'ga; (XIII) Tui–Barabinsk; (XIV) Kolyvan'–Tomsk; (XV) Minusinsk–Salair; (XVI) Taimyr; (XVII) Central Siberian; (XVIII) Rybinsk; (XIX) Tuva; (XX) island arcs; (XXI) Kazakhstan and Angarida land masses. Paleolandscape regions: (XIa) Vakh; (XIb) Taimyr; (XIIIa) Barabinsk; (XIIIb) Tui.

The Khanty-Mansiisk paleolandscape occupied a small area adjacent in the north to the Kustanai–Tyumen paleolandscape. It corresponded to a sea gulf having northern connections with the Shaim sea (Fig. 2) and extending inland for a distance of about 250 km within the northern alluvial denudation plain of the Kazakhstan continent. The gulf was in contact with the Kustanai–Tyumen paleolandscape in the west and with the Kazakhstan and Dem'yanka paleolandscapes in the east. Hydrodynamics in the gulf was calm. Conodonts

Palmatolepis sp. and *Polygnathus* sp., the inhabitants of the gulf, coexisted with rare foraminifers during the early Givetian time.

The Kazakhstan paleolandscape, like that of the Kustanai–Tyumen area, represented the vast alluvial–diluvial plain of the corresponding, formerly uplifted continent. Red silty and clayey sediments characteristic of the plain were deposited under condition of arid climate, when long droughty periods alternated with short humid and rainy seasons that favored development of vegetation, especially in southern areas. Among plant remains buried in sediments, there were distinguished *Taeniocrada dubia* Kr. et W., *Artschaliphyton inicum* Senk., *Lepidodendropsis kazachstanica* Sok., *Lidasimophyton akkerrmensis* Senk., *Protolepidodendron scharianum* Kr., *Hyenia* cf. *sphenophylloides* Nath., and *Protopteridium hostimense* Kr. (Dubatolov *et al.*, 1977; Senkevich, 1967; 1991). In northern areas of the alluvial–diluvial plain, there was the Dem'yanka paleolandscape that accumulated shingles and siliceous clay.

The Tevriz paleolandscape that bounded the Kazakhstan and Dem'yanka paleolandscapes in the east corresponded to another part of the alluvial–diluvial plain. Clayey and sandy deposits dominating here suggest that seawaters periodically flooded this area of the plain. The inevitable frequent mixing of fresh and marine waters apparently created environments unfavorable for both the terrestrial vegetation and marine fauna.

The Nyurol'ka–Var'egan paleolandscape represented a narrow marine basin that bordered the alluvial–diluvial plain of Kazakhstan and extended from southernmost areas of the West Siberian plain up to the latitude of Nadym. In contrast to adjacent paleolandscapes, the gently undulating bottom of the basin was covered with calcareous ooze, clay, sand, and acidic lavas. The shallow shelf zone of the basin was populated by foraminifers *Bykovaella uralica* (L. Petr.), *B. bykovae* (Pojark.), *Salpingothurammina tuberculata* (Lip.), *S. gekkeri* (Antr.), *Neoivanovella simplexituba* Tchuv. et Juf., *Parathurammina graciosa* Pron., *Auroria sphaerica* Sab., *Suleimanovella paulis* (E. Byk.), *S. elegans* (Pojark.), and others, and also by ostracods *Bairdia kynovensis* Rozhd., *Samarella crassa* (Pojark.), and *Orthocypris subparallella* Pol. The coexisting conodont species were *Polygnathus decorsus* (Stauff.), *Icriodus expansus* Brans. et Mehl, *I. symmetricus* Brans. et Mehl, *I. alternatus* Brans. et Mehl, *Belodella* aff. *serrata* Khod. et Tschern. Tabulate corals were represented by 21 species, dominant among which were *Favosites goldfussi* Orb., *F. tuimazaensis* Sok., *Pachyfavosites polymorphus* (Goldf.), *P. alpensis* (Winch.), *Emmonsia aspera* Yanet, *Thamnopora cervicornis* (Blainv.), *T. certa* (Tchern.), *Gracilopora vermicularis* (McCoy), *Dendropora dubrovensis* Dubat., *Alveolitella fecunda* (Salee sensu Lec.), *A. polenowi* (Peetz), *Scoliopora denticulate* (M.-Edw. et H.). Tentaculitid species *Viriatellina hollardi* Lard. were relatively rare. Within the coral–stromatoporoid pavement, massive and ramose stromatoporoids *Actinostroma salairica* Yavor., *Amphipora ramosa ramosa* (Phil.), *A. ramosa desquamata* Lec., and *Stellopora laxeperforata* (Lec.) were tightly associated with tabulate corals. More than 15 acritarch species, the dominant *Crassisphaeridium inustatum* Wicander, *Multiplicisphaeridium andrewsi* (St. et Will.) Eisenack, Cramer et Diez, *M. grosjeani* (St. et Will.) Eisenack, Cramer et Diez, *M. spicatum* Staplin, and *Lophosphaendium ochthos* Wicander included, populated pelagic settings.

Population of the relatively deep shelf included the following fauna groups: tentaculitids *Nowakia autentalensis* Alb. and *N. sulcata* (Roem.); conodonts *Polygnathus varcus* Stauff., *P. angusticostatus* Witt., *P. disparilis* Ziegl., Klapp. et Joung., *P. xylus ensensis* Ziegl., Klapp. et Joung., *Icriodus difficilis* Ziegl. et Klapp., *I. brews* Stauff., *I. subterminus* Brans., *I. expansus* Brans, et Mehl, *I. obliquimarginatus* Bisch. et Ziegl., *I. symmetricus* Brans, et Mehl, *Palmatolepis disparilis* Ziegl., Klapp. et Joung., *Ozarcodna brevis* (Bisch. et Ziegl.); ostracods *Samarella crassa* Pol., *Bairdia kynovensis* Rozhd., *B. aperta* Pol., *Cavellina accurate* Pol., and *Orthocypris subparallela* (Pol.). They associated with rare foraminifers *Parathurammina irregularis sibirica* Saev, and *P. graciosa* Pron., and also with worms *Polychaetaspis* cf. *wyszogrodensis* Kozlowski (*Paleozoi yugo-vostoka...*, 1984; *Stratigrafiya paleozoiskikh...*, 1990).

The Vakh–Nadym paleolandscape was adjacent to the northeastern Nyurol'ka–Var'egan paleolandscape and represented a marine basin, where the bottom consisted of shallow gullies and gentle rises. This paleolandscape is subdivided into two regions.

The Vakh region near the Nyurol'ka–Var'egan paleolandscape (Fig. 2) had either the calcareous–clayey–sandy, or tuffaceous–silty substratum at the bottom. It is likely that the region was bounded by two submeridional faults slightly turning north-northwest within the northern sector and thus following the eastern boundary of the alluvial–diluvial plain of Kazakhstan. Data on fossil remains from this region are unknown.

The Nadym paleolandscape region was situated to the east and extended as a narrow belt bounded by two faults of the north-northeastern strike. In contrast to the Vakh region, it was devoid of calcareous ooze and hosted eruptions of basic lavas. Data on inhabitants of this basin are also unknown. Judging from the composition of sediments consisting of clayey-sandy and tuffaceous material, the basin was not deeper than few dozens of meters.

The Sil'ga paleolandscape was in the middle of four other paleogeographic areas: the Nyurol'ka-Var'egan paleolandscape in the west, Nadym region in the east, Vakh region in the north, and Tui region in the south. The water mass in this quiet basin was of normal salinity and fairly well illuminated. The terrigenous influx

was insignificant, and calcareous ooze covered the entire bottom area in the Sil'ga basin. In accordance with habitat environments, the basin was populated by communities of foraminifers, stromatoporoids, ramose corals, bryozoans, and conodonts. Foraminifers *Parathurammina crassitheca* Antrop., *Cribrosphaeroides* aff. *ovalis* Pojark., and *Neoarchaesphaera polypora* (Antrop.) migrated in the basin from the Tui–Barabinsk and Kolyvan'–Tomsk paleolandscapes.

The Tui–Barabinsk paleolandscape comprising two respective regions corresponded to a narrow basin elongated in the meridional direction. The Nyurol'ka-Var'egan and Sil'ga paleolandscapes bordered it in the north. The Barabinsk region accumulated clayey–calcareous ooze, siliceous clay, fine-grained sand, and tuffaceous material. The terrigenous clayey–sandy material was derived from the alluvial–diluvial plain. The basin could be as deep as 100 m. Remarkable inhabitants of pelagic settings in the basin were peculiar radiolarians representing families Staurosphaeridae and Stilosphaeridae (Bogush *et al.*, 1975). The Tui paleolandscape region was situated eastward, and this narrow basin extended from south to the north up to the boundary with the Sil'ga region (Fig. 2). Siliceous clay represented the main type of sediments in the basin that was populated by rare radiolarians *Staurostylus* sp. and *Stylostaurus* sp.

The Kolyvan'–Tomsk paleolandscape comprising areas of the Novosibirsk and Kemerovo districts was located to the southeast of the Tui-Barabinsk basin. Gentle hills and shallow gullies at the bottom of the Kolyvan'–Tomsk basin accumulated various types of deposits: calcareous ooze (15%); clay (15%); sand (20%); tuffaceous sand (20%), and intermediate to basic lavas (30%). Being inside the arid zone, the basin was warm (water temperature not lower than 18–20°C) and had normal salinity thus attracting benthic immigrants, which populated settings above the depth range of 60–100 m. Principal members of benthic population were ramose and massive corals associated with brachiopods. The most widespread were coral pavements composed by tabulate forms *Pachyfavosites polymorphus* (Goldf.), *Thamnopora angusta* Lec., *Crassialveolites* aff. *cavernosus* (Lec.), *Tyrganolites* sp. Heliolitids *Heliolites vulgaris* Tchern. were less abundant, and the assemblage of rugose corals included *Grypophyllum gracile* Wdkd., *Campophyllum litvinovitschae* Soshk., *Lythophyllum divisium* Wdkd., and *Tabulophyllum* sp. Species *Euryspirifer cheehiel* (Kon.), *Chonetes coronata* Conr., and *Atrypa* ex gr. *reticularis* L. represented brachiopods (Besprozvannykh, 1964). It should be mentioned that the Kolyvan'–Tomsk paleolandscape was the only one volcanically active in the Altai– Sayan territory and hosted submarine eruptions of intermediate to basic lavas.

The Minusinsk–Salair paleolandscape corresponded to a vast sea basin that originated when seawaters flooded many areas of southern Siberia between the ore-bearing and southern districts of the Altai Mountains in the west to the Minusinsk depression in the east, the piedmonts of the eastern Sayan Mountains included. This sea transgression was greatest one during the Devonian period. It influenced not only the sedimentation environments, but also the dispersal of marine benthos. Arriving in southern Siberia from Mongolia and China, the transgression reached the Kolyvan'–Tomsk regions, but it did not penetrate further northward in the West Siberian plain. At least, the late Givetian species *Euryspirifer cheechiel* (Kon.) widespread in southern Siberia did not penetrate that far. Apparently, transgressions periodically advanced into the West Siberian sea from different sides, and the sea represented a kind of a transit basin. In different periods, many groups of species migrated through the sea from paleobasins of the East European platform, Urals, Taimyr, northeastern Russia, Siberian platform, Mongolia, China, Kazakhstan, and Central Asia. Thus, the West Siberian plain is a key area for various paleogeographic reconstructions and also for general understanding of geological evolution of Siberia as whole. Accordingly, the West Siberian sea cannot be considered as an independent paleobiogeographic province comparable with the Uralian and Altai–Sayan provinces.

The Givetian transgression is recorded in sedimentary sequences of the Taimyr and Central Siberian seas that occupied considerable areas. In the center of southern region of the West Siberian sea, seawater flooded the marginal zone of the Kazakhstan continent. As a result, the alluvial denudation plain appeared in the north of the continent and accumulated coarse terrigenous material that was derived from the inner uplifted areas of the Kazakhstan land mass. The transgression was responsible for a certain humidification of the arid climate and created conditions favorable for development of terrestrial plants, especially of lycopsids (Dubatolov *et al.*, 1977; Senkevich, 1991), but it did not influenced the Angarida continent. The latter even grew in size advancing from the present-day valley of the Yenisei River into eastern areas of the West Siberian plain.

Salinity in the Minusinsk–Salair sea was normal in general and slightly elevated in its eastern zone, as one can judge from gypsum strata present in the Northern Minusinsk depression. An intense faunal interchange and dispersal of closely related species communities was characteristic of areas corresponding to the present-day Minusinsk depression and northeastern region of the Kuznetsk basin. The Minusinsk–Salair sea presumably was shallow, having the gently undulating bottom relief. This is evident from abundant algae remains forming, for instance, thick beds of algal limestone in the Beiskoe Horizon of the Minusinsk depression, and also from widespread occurrence of coral–algal bioherms and reefs.

In the late Givetian seas, algae periodically were principal dwellers associated with ramose and massive corals, brachiopods, crinoids, and ostracods. Bivalves, foraminifers, bryozoans, trilobites, conodonts, and phyllopods were less abundant. It is important to note diversification of thermophilic organisms at that time, primarily of tabulate and rugose corals (about 80 and 40 species, respectively), bryozoans (more than 30 species), and brachiopods (more than 100 species). Dominant among tabulate corals were *Pachyfavosites polymorphus* (Goldf.), *Neoroemeria westsibirica* Rad., *N. soshkinae* Tchud., *Dendropora dubrovensis* Dubat., *Thamnopora nicholsoni* (Frech), *T. angusta* Lec., *Alveolites mallieuxi* Lec., *Alveolitella polenowi* (Peetz), *Crassialveolites cavernosus* (Lec.), *C. crassiformis* (Sok.), *Placocoenites escharoides* (Stein.), *Tyrganolites eugeni* Tchern., *Natallophyllum giveticum* Rad., and *Scoliopora denticulate* (M.-Edw. et H.) (Dubatolov, 1959, 1963). Principal species of rugose corals were *Atelophyllum cyindricum* (Schlüt), *Neostringophyllum waltheri* (Joh), *Dendrostella rhenana* (Frech), *Columnaria vagranensis* Soshk., and *Grypophyllum gracile* Wdkd. (Ivaniya, 1965). Bryozoan assemblage mostly consisted of *Neotrematopora typica* Moroz., *Fistulipora changi* (Jang.), *Lioclema heitalensis* Jang., *L. yakovlevi* (Shoen.), *L. celebratum* Moroz., *Fenestrella vera* Ulr., and *Hemitrypa devonica* Nekh. In brachiopod communities, there were abundant *Gypidula planispinosa* Grabau, *Schellwinella umbraculum* (Schl.), *Productella subaculeata* (Murch.), *P. productoides* (Murch.), *Uncinulus parallelepipedus* (Bronn), *Camarotoechioides lazutkini* (Ržon.), *Desquamatia desquamata* (Sow.), *Undispirifer undiferus* (Roem.), *Indospirifer pseudowilliamsi* Ržon., *Euryspirifer cheehiel* (Kon.), *Mucrospirifer mesocostalis* (Hall), *Reticulariopsis aviceps* (Kays.), *Cyrtina heteroclita* Defr., *Retzia lopatini* Stuck., *Stringocephalus* ex gr. *burtini* Defr., *S. sibirica* E. Ivan., *Denckmanella damesi* (Holz.), and *Spinocyrtia martianofi* (Stuck.). In addition, there were the following faunal groups: foraminifers *Cribrosphaeroides simplex* (Reitl.) and *Parathurammina graciosa* Pron. (Zadorozhnyi, 1987); crinoids *Hexacrinus kartzevae* Yelt. and *Sphaerocrinus geometricus* Goldf.; trilobites *Proetus lazutkini* N. Tchern., *P. peetzi* N. Tchern., and *Scutellum tenuistriatum* (N. Tchern.); ostracods *Moorites lebedjanicus* Pol., *M. legibilis* Pol., *Bairdia safonovskensis* Pol., *B. summacuminata* Cor. et Malk., and *Janetina prima* Pol.; phyllopods *Estheria membranacea* (Pacht) and *E.* ex gr. *pogrebovi* Lutk.; cephalopods *Gomphoceras* (?) *poculum* Hall, *Agoniatites fulguralis* (Whitb.), *A. coptulata* Arch. et Vern., *A. obliquus* Whitb., and *Sellanarcestes discoids* (Wald.) (Dubatolov, 1959, 1963; Chudinova, 1964; Rzhonsnitskaya, 1952, 1968, 1975; Morozova, 1960; Polenova, 1960; Dubatolova, 1971, 1972, 1977; Ivanova *et al.*, 1964; Ivaniya *et al.*, 1974; Chuvashov *et al.*, 1985).

Despite the nonmarine origin, the terrestrial sediments of the paleolandscape were interrelated, in our opinion, with marine deposits, because they reveal ripple marks, exsiccation cracks, and impressions of rain drops usually characterizing the coastal sedimentation settings. These sediments yield remains of plants *Psillophyton salairicum* An. et Step., *Pectinophyton norvegicum* Höeg., *Protocephalopteris praecox* (Höeg.) An., *Taeniocrada gracilis* Tchirk., *Svaibardia polymorpha* Höeg., *Aneurophyton germanicum* Kr. et Weyl., and *Barsassia ornata* Zal. (Rzhonsnitskaya *et al.*, 1985; Stepanov, 1975).

The considered diversity of benthic organisms suggests favorable habitat environments in the basin: normal salinity (Dubatolova, 1973), warm climate (20–22°C), fairly calm hydrodynamics, depth range from several to 40–60 m, perfect illumination of sea bottom, and free connections with other basins.

The Taimyr paleolandscape adjacent to the northeastern periphery of the West Siberian sea is characterized by scarce data mostly concerning its southern areas. These areas predominantly accumulated calcareous and dolomitic ooze. The Taimyr basin was populated by brachiopods, ostracods, crinoids, stromatoporoids, and algae. Two latter groups growing in algal–stromatoporoid meadows often formed biohermal massifs. The species diversity of benthic faunas was moderate. *Stringocephalus* sp., *Chascothyris* sp., *Undispirifer undiferus* (Roem.), *Uncinulus parallelepipedus* (Bronn), and *Gypidula biplicata* Schnur represented brachiopods, and the genus *Cupressocrinus* characterized crinoids (*Resheniya...*, 1982).

The Central Siberian paleolandscape was situated in the northwest of the present-day Siberian platform. An intense accumulation of calcareous ooze was a characteristic feature of this basin except for its area adjacent to the northwestern platform margin, where sedimentation of dolomitic ooze was in progress. This area was populated by foraminifers, ramose and massive corals, brachiopods, trilobites, and conodonts. Remarkable species were as follows: *Moravammina fragilis* Byk., *M. segmentata* Pok., *M. plena* Reitl., and *Proninella tamarae* Reitl. among foraminifers; *Pachyfavosites polymorphus* (Goldf.), *Thamnopora proba* Dubat. and *T. cervicornis* (Blainv.) among tabulate corals; *Hexagonaria* ex gr. *baschkirica* (Spassk.), *Embolophyllum longiseptatum* (Bulv.), and *Neostringophyllum* cf. *waltheri* Smith among rugose corals; *Productella djaltulensis* Nal., *Striatoproductella tungusensis* (Nal.), *Atrypa zonata* Schnur., *Desquamatia desquamate* (Sow.), *Spinatrypa aspera* (Schloth.), *Schizophoria striatula* Schloth., *Emanuella takwanensis* (Kays.), *E. subumbona* (Hall), *Undispirifer undiferus* (Roem.), *Athyris concentrica* Buch, and *A. globosa angeliciformis* Mark. among brachiopods; *Polygnathus linguiformis gamma* Bultynck, and *Panderodus unicostatus* Brans et Mehl among conodonts (Dubatolov, 1978; Krylova and Menner, 1962; Menner, 1972; Matukhin *et al.*, 1995).

The Rybinsk paleolandscape corresponded to a small area of the southwestern Angarida adjacent to the

northeastern periphery of the Minusinsk–Salair sea. Variegated silty to sandy alluvial deposits represented the main type of sediments deposited in that area. The denuded Angarida land mass was a provenance of terrigenous material. The alluvial Rybinsk plain was periodically flooded by fresh waters and hosted basins that were populated by phyllopods *Ctenaenigma tuvaense* Nov. Coastal plants around the basins were represented by *Uralia minussinskiensis* Petr., *Tomiphyton primaevum angusta* Petr., and *Protobarinophyton obrutschevi* An.

The Tuva paleolandscape corresponded, like the previous one, to a small alluvial plain that opened southward into the Mongolian sea and was bordered in the north by the hilly denudation plain of Angarida. Shallow fresh-water basins of the paleolandscape predominantly accumulated red clay, silt, and sand. Phyllopods *Pseudestheria pogrebovi* (Lutk.), *Ulugkemia barikensis* Nov., *U. minussensis* Nov., *Glyptasmussia membranacea* Pacht., *Trigonestheria altaikensis* Nov., and some others inhabited these basins. The assemblage of coastal plants consisted of *Aneurophyton germanicum* Kr. et Weyl., *Protopteridium* sp., *Rhacophyton* sp., *Glyptophyton granulare* Krysht, and *Hyenia* sp. Variegated sediments with similar fossils were also accumulated in shallow, occasionally desiccating basins of the Minusinsk depression. The basins were of the intracontinental lacustrine type.

The above data characterizing the late Givetian transgression into the western and central Siberia show that faunas of the West Siberian and Minusinsk–Salair seas were more diverse than those from basins of central Siberia. Many species typical of the latter were of a wide geographic range and migrated far away from Siberian seas into other biogeographic provinces. In addition to the West Siberian, Minusinsk–Salair, Taimyr, and Central Siberian seas, species of this kind populated basins of the European platform, Indigirka–Kolyma province, China, and Kazakhstan. Thus, the Givetian Age was indeed a time of destruction of many geographic barriers that enabled an intense migration of various faunas and development of many cosmopolitan species (Dubatolov, 1972, 1984). As compared to the Eifelian time, bioherms and even biohermal and reefal massifs became more abundant, especially during the late Givetian period. Biohermal massifs composed of rugose corals are widespread in the Mongolian Altai and Minusinsk depression. In the Salair Mountains, the coral–stromatoporoid–algal bioherms are typical of the Safronovo Horizon, and biohermal or reefal massifs built by tabulate and rugose corals are abundant in the Alchedat Horizon. Similar biogenic buildups that originated in the Nyurol'ka basin of the West Siberian plain are now concentrated in the upper portion of the Gerasimov Formation (Bol'shakova *et al.*, 1994; Ivanova *et al.*, 1964; Krasnov and Stepanov, 1979; Krasnov *et al.*, 1980, 1986; Sharkova, 1980; Stepanov, 1970; Stepanov *et al.*, 1984, 1990).

The prevalence of marine sedimentation settings and habitat environments was also characteristic of the Frasnian Age. However, this age was distinct because of cessation of volcanic and tectonic activity in western regions of the West Siberian sea. This activity that became lower than before was concentrated in central areas of the sea.

A land mass corresponding to the northern part of the Kazakhstan continent appeared in the south of the West Siberian sea. It was smaller than in previous epochs and did not extend northward further than the middle courses of the Irtysh River. It is likely that the land area of Angarida was reduced as well. The western Angarida was flooded by sea, and its land areas advanced into the region corresponding to present-day valleys of the Podkamennaya Tuguska and Angara rivers. In contrast, the Kazakhstan continental area grew, and the Irtysh–Vakh and Ob' land masses arose in the central zone of the West Siberian sea. The colossal sea basin shoaled in general, and differentiation of paleolandscapes became less prominent than before. We distinguished 15 paleolandscapes of the Frasnian time (Fig. 3).

The Cis-Urals paleolandscape that slightly grew in size retained its previous contours, in general. It accumulated predominantly terrigenous deposits, among which the share of calcareous ooze was about 20%, not more. Andesite lavas represent here only 2% of the rock sequence. The thick-walled brachiopods and foraminifers populated the respective basin, where hydrodynamics was active. The bottom relief apparently was undulating, and the depth was not greater than 10–20 m. The influx of fresh underground and surface waters frequently disturbed the basin salinity, and population of the paleolandscape was rather poor. Species *Archaesphaera magna* Sul., *Parathurammina* sp., *Nanicella uralica* Tchuv., and *Tikhinella multiformis* (Lip.) characterize the local assemblage of foraminifers, and *Cyrtospirifer conoideus* (Roem.) is the known representative of brachiopods.

The East Uralian paleolandscape was separated from the Cis-Urals basin by a rise and accumulated silty and calcareous sediments. Limited data of deep drilling suggest that the paleolandscape under consideration was populated by foraminifers, ramose stromatoporoids, tabulate corals, brachiopods, and algae. The corresponding basin was quiet in hydrodynamic aspect and sheltered from provenances that supplied the Cis-Urals paleolandscape with terrigenous material. The basin was up to 60 m deep, and its water mass was aerated to a sufficient extent. Quiet conditions of sedimentation resulted in a fine horizontal bedding of deposits. The following fossil groups exemplify population of the basin: foraminifers *Parathuramminites obnata* (Tchuv.), *P.* sp., *Parathurammina breviradiosa* Reitl., *Irregularina karlensis* Viss., *Uralinella bicamerata* E. Byk., *Radiosphaera basilica* Reitl., *Vicinesphaera squalida* Antrop., *Bisphaera elegans* Viss., and

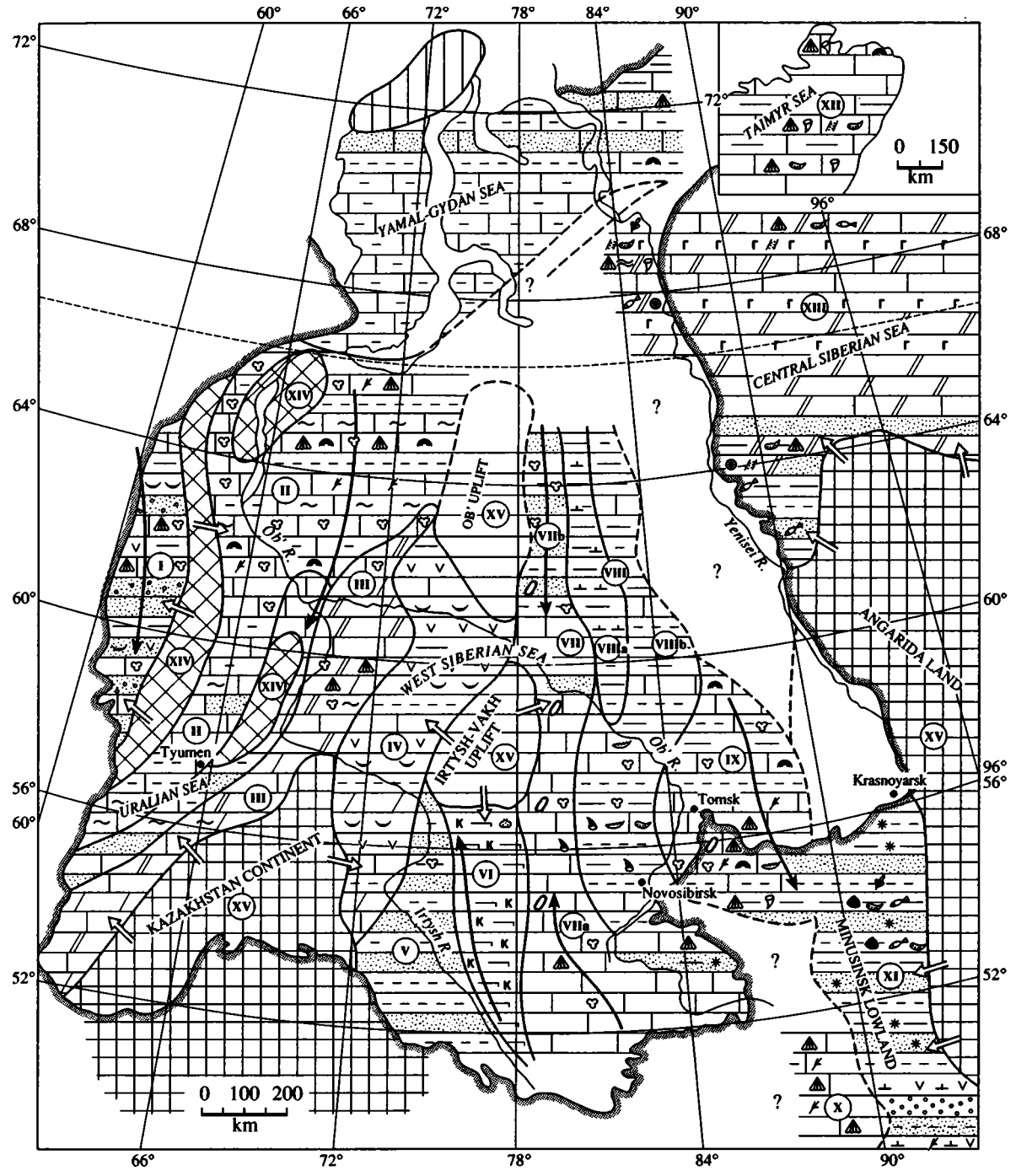

Fig. 3. Scheme of Frasnian paleolandscapes in Siberia (symbols as in Fig. 1).
Paleolandscapes: (I) Cis-Urals; (II) East Uralian; (III) Kustanai–Tyumen; (IV) Salym; (V) Tevriz; (VI) Tui–Barabinsk; (VII) Nyurol'ka–Var'egan; (VIII) Nikol'skii; (IX) Kolyvan'–Tomsk; (X) Altai; (XI) Yenisei; (XII) Taimyr; (XIII) Central Siberian; (XIV) island arcs; (XV) Kazakhstan and Angarida land masses, and Ob' and Irtysh–Vakh uplifts. Paleolandscape regions: (VIIa) Nyurol'ka; (VIIb) Var'egan; (XIIIa) Vakh; (XIIIb) Tym.

Baituganella ferganensis Pojark.; stromatoporoids *Amphipora* cf. *laxeperforata* Lec. and others; tentaculitids *Viriatellina devoniana* G. Ljasch.; tabulate corals *Crassialveolites* ex gr. *parvus* (Lec.) and *Alveolitella* sp.

The easterly narrow zone of the Kustanai–Tyumen paleolandscape extended from the southwest to the northeast widening in the middle. The bottom substratum in the corresponding basin consisted mostly of dolomitic and marly sediments; calcareous ooze was

less widespread (about 10%). Lithology of sediments is indicative of arid climate and shallow sedimentation settings in a moderately aerated and reasonably warm basin. Foraminifer and brachiopod species from the basin suggest that it was well connected with the East Uralian basin, although there was a narrow and low island striking northeast and partially separating the basins (Fig. 3). The Kustanai–Tyumen paleolandscape was situated between the East Uralian basin in the west, Kazakhstan continent in the southeast, and volcanically active Salym paleolandscape in the east. This geographic situation presumably was responsible for the calcareous–marly–dolomitic type of sedimentation. To some extent, the character of sedimentation is indicative of an elevated water salinity within the basin populated by rare representatives of foraminifers and brachiopods.

It should be noted that there was an important feature of benthos distribution in the Cis-Urals, East Uralian, and Kustanai–Tyumen paleolandscapes of the West Siberian sea and in the Uralian sea located southwestward of the latter. Only a dozen of foraminifer species and few representatives of brachiopods, tabulate corals, and tentaculitids are known from the listed paleolandscapes, whereas fossil assemblages from the Uralian sea include more than 30 foraminifer taxa (Chuvashov, 1968; Zadorozhnyi, 1987), dozens of rugose coral forms (Soshkina, 1949; Spasskii, 1977), abundant brachiopods (*Devon Zapadnogo...*, 1952), and remains of many other organisms.

The Salym paleolandscape was surrounded by the following paleogeographic areas: the Kustanai–Tyumen paleolandscape in the west, the Kazakhstan continent in the southwest, the Ob' uplift in the northeast, the Charysh uplift in the southeast, and the Tevriz paleolandscape in the extreme southeast (Fig. 3). The volcanically active Salym basin accumulated intermediate lavas and their tuffs against the background of deposition of silt and calcareous to marly ooze. The active tectonic environments and proximity of the basin to the Kazakhstan, Ob', and Irtysh–Vakh land masses controlled the shallow state and active hydrodynamics of the basin, and also the accumulation of silty material, suspension of which negatively affected the water illumination. The insufficient illumination of the bottom and the salinity decrease were apparently unfavorable, like some other ecological factors as well, for the proper development of marine benthic organisms. Only the northern area of the basin was populated by scarce foraminifers *Cribrosphaeroides* sp. and *Tscherdyncevella* sp., and also by algae *Girvanella* sp. (Bochkarev and Pogorelov, 1973).

The Tevriz paleolandscape located southeastward of the Salym basin was in proximity to the Kazakhstan continent in the west and to the Irtysh–Vakh land mass in the north (Fig. 3). Accordingly, silty and sandy deposits were dominant in the paleolandscape that represented a shallow marine basin with terrigenous sedimentation compensating the bottom downwarping. The slightly freshened water of the basin had temperature abut 18–20°C. Sparse benthic population included foraminifers *Nanicella* aff. *porrecta* E. Byk. and *Nodosaria* sp. (*Problemy yarusnogo...*, 1984), the presumable immigrants from the Kurgan region (Bogush *et al.*, 1975).

The Tui–Barabinsk paleolandscape corresponded to a narrow zone between the Tevriz and Nyurol'ka–Var'egan paleolandscapes located respectively in the west and in the east. The basin in that zone was rather deep and accumulated siliceous clay.

Thus, going from the west to the east, we can see that the accumulation of calcareous, marly, sandy and volcanogenic material (tuffs and lavas of intermediate composition) in the Salym basin was replaced first by the terrigenous sedimentation in the Tevriz basin, and then by deposition of fine-grained clayey–siliceous material in the Tui–Barabinsk though. The trough was relatively deep, bordered by two faults, and represented, as one could think, an "incipient oceanic basin" populated by radiolarians, the dwellers of deep waters. This assumption is however incompatible with data of Osyko (1958), who identified miospores *Zonotriletes acucularis* Andr., *Z. maculates* Andr., *Z. argutalus* Andr., *Z. armigerus* Andr., and *Z. alloplectus* Andr. in brown shales recovered by the reference Barabinsk borehole. As there is a chance that the above miospores were transported in the basin from the nearby land, the problem of "incipient ocean" is still a subject for consideration. On the other hand, coeval deposits in the Verkhnyaya Dem'yanka area yielded foraminifers *Parathurammina dagmarae* Sul., *P. paulis* (E. Byk.), *P.* cf. *subvesta* E. Byk., *Cribrosphaeroides* sp., *Archaelagena mirabilis* Pojark., *A. shesmae* Antrop., *Neoarchaesphaera polypora* (Antrop.), *N. radiata* M. Maclay, and *N.* aff. *magna* (Sul.) (Bogush *et al.*, 1975). Here, we should keep in mind that some of the listed species are of a wide stratigraphic range spanning the interval of the Givetian to Famennian Stage inclusive. Simultaneously, they are important indicators of the basin depth, and, for instance, *Parathurammina paulis* (E. Byk.) inhabited settings at the depth of about 10 m (Poyarkov, 1969, 1979).

The Nyurol'ka–Var'egan paleolandscape extended as a narrow submeridional zone separating the Tui–Barabinsk basin and Irtysh–Vakh uplift from the Nikol'skii paleolandscape located eastward. The zone is subdivided into the Nyurol'ka and Var'egan regions.

The Nyurol'ka region in the southern wider part of the zone predominantly accumulated calcareous ooze, and its benthic population included foraminifers, stromatoporoids, massive to ramose tabulate corals, tentaculitids, brachiopods and conodonts. The following species groups worth to be mentioned here: *Salpingothurammina* sp., *Neoarchaesphaera polypora* (Antrop.), *Eonodosaria evlanensis* Lip., *Multiseptida corallina* E. Byk., *Archaesphaera minima*, Sul., *A. suleimanovi*

Bog. et Juf., and *A. crassa* Lip. among foraminifers; *Stachyodes singularis* Yavor. and *Amphipora laxeperforata* Lec. among stromatoporoids; *Emmonsia aspera* Yanet., *Thamnopora boloniensis* (Goss.), *T. polyforata* (Schloth.), *Gracilopora vermiculata* (McCoy), *Crassialveolites domrachevi* (Sok.), and *Trypanopora* sp. among tabulate corals; *Homoctenus kikinensis* G. Ljasch. and *Styliolina devoniana* G. Ljash. among tentaculitids; *Polygnathus xylus* Stauffer, *P. decorosus* Stauffer, *P. asymmetricus ovalis* Ziegler et Klapper, *P. dubius* Hinde, *P.* cf. *normalis* Branson et Mehl, *P. communis communis* Branson et Mehl, *Icriodus difficilis* Ziegler, Klapper et John., *I. alternatus* Branson et Mehl, *I. brevis* Stauffer, *Ancyrodella rotundiloba bindosa* Uyeno, *Palmatolepis hassi* Müller et Müller, *P. subrecta millerex* Young., *P. foliacea* Young., and *Ancyrognathus triangularis* Young among conodonts. Chitinozoans, calcisphaerids, and acritarchs inhabited pelagic settings. Algae were represented by *Renalcis devonicus* Johnson, *Shuguria flabeliformis* Antr., *Izhella nubiformis* Antr., *Rectagulina tortuosa* (Antr.), *Epiphyton buldyricum* Antr., *Kamaena* cf. *tatarstanica* (Antrop.), *Kamaena* sp., *Girvanella ducii* Wether, *G. problematica* Nich., *Rothpletzella devonica* (Masl.), *Polyderma chovanensis* Reitl., and by associated charophytes *Umbella bella* Masl. and *Quasiumbella nana* (Reitl) (*Paleozoi yugo-vostoka...*, 1984; Chuvashov *et al.*, 1985; Zadorozhnyi, 1987; *Stratigrafiya paleozoiskikh...*, 1990). Many benthic organisms lived on pelagic chitinozoans, calcisphaerids, and acritarchs.

The fine parallel bedding of calcareous deposits suggests quiet hydrodynamic environments of sedimentation. The muddy substratum was populated by fossorial organisms that burrowed ooze in various directions relative to the bedding planes. Abundance of ramose stromatoporoids and tabulate corals is also indicative of the inactive hydrodynamics. The basin was rather shallow, not deeper than 60–80 m. The water salinity apparently was close to normal, as it is evident from frequent occurrence of stenohaline species in the fossil assemblages, especially among stromatoporoids and tabulate corals. At the Frasnian time, similar conditions were characteristic of the Uralian sea (Chuvashov, 1963).

The northern Var'egan region extending, like the previous one, from the south to the north was bounded in the east by the Nikol'skii paleolandscape. Deposits of the region mostly consisted of calcareous ooze and fine-grained sand. The terrigenous material of the latter was derived from the hilly Ob' uplift located in the west. In addition, there were algal reefs in the region; they represented a continuation of reefal buildups located in the Nyurol'ka region. Hydrodynamics could be fairly active, and salinity of the basin was controlled by the fresh-water influx from the side of the Ob' uplift. These factors presumably decelerated the development of benthic fauna, the known representatives of which are foraminifers *Parathurammina* sp. and *Cribrosphaeroides* sp. The basin depth was not greater than 60 m.

The relatively wide zone of the Nikol'skii paleolandscape located eastward of the previous one is subdivided into the Vakh and Tym regions. The Vakh region adjacent to the Nyurol'ka–Var'egan paleolandscape accumulated clay and subordinate flows of basaltic lavas. As we think, this region corresponded to the deepest area of the West Siberian sea. Fossils from this area are unknown as yet, and, accordingly, we have no guide to estimate salinity and temperature in the basin. Apparently, these parameters were here close to those in the Nyurol'ka–Var'egan paleolandscape. The boundary between the Vakh and Tym regions is conditional. The latter was adjacent to the Kolyvan'–Tomsk paleolandscape in the southeast and accumulated, in contrast to the former, calcareous ooze, clay, and silty material. The Tym region devoid of volcanic activity was populated by foraminifers, tentaculitids, ostracods, and conodonts. Species *Parathurammina crassitheca* Antrop., *Cribrosphaeroides* aff. *ovalis* Pojark., and *Neoarchaesphaera polypora* (Antrop.) represented the foraminiferal assemblage (*Problemy yarusnogo...*, 1984; Bogush *et al.*, 1975). The known tentaculitid taxa are *Homoctenus acutus* Ljasch., *Crassilina timanica* Ljasch., *Viriatella elongata* Ljasch., *Styliolina gassanovae* Ljasch., and *S. domaniceuse* Ljasch. (Meretskaya, 1993). The ostracod assemblage included *Famenella kairovensis* Rozhd., *Eonodosaria evlanensis* Lip., and some other species (Savina, 1993). Conodonts were exemplified by *Polygnathus asymmetricus ovalis* Zieg. et Klapp., *P. dubius* Hinde., *P.* cf. *latifossatus* Wirth., and *Ancyrognathus triangularis* Young. The basin could be as deep as 60 m (Poyarkov, 1969). As compared to the Vakh region, hydrodynamics here was less active, because sediments retain intact the horizontal bedding. The water temperature was about 18–20°C.

The Kolyvan'–Tomsk paleolandscape to the southeast of the Tym region predominantly accumulated calcareous ooze; sand and silt were concentrated in the local Gorlovo trough. Benthic population included foraminifers, stromatoporoids, tabulate and rugose corals, bryozoans, algae, and brachiopods; the latter represented the main group of organisms in the Gorlovo trough. Characteristic of foraminifers were *Salpingothurammina tuberculata* (Lip.), *Suleimanovella paulis* (E. Byk.), *S. suleimanovi* (Lip.), *Bisphaera malevkensis* Bir., *Paratikhinella cannula* E. Byk., *Nanicella bella* E. Byk., *N. porrecta* E. Byk., *N. tchernyshevae* Lip., *Tikhinella multiformis* (Lip.), *Eotuberitina crassa* Pojark., *Neoarchaesphaera polypora* (Antrop.), *Eonodosaria evlanensis* Lip., and *Eogeinitzina devonica* Lip. (Zadorozhnyi, 1987). *Amphipora* sp. represented stromatoporoids; *Thamnopora boloniensis* (Goss.), *Alveolites suborbicularis* Lam., *Crassialveolites obtortus* (Lec.), and *Thecostegites bouchardi* (Mich.) were principal forms of tabulate corals, and *Peneckiella isylica* Bilv. together with *Megaphyllum bulvankeri* Soshk. exemplified the assemblage of rugose corals (Ivaniya,

1965). The listed taxa coexisted with bryozoans *Amplexopora kisilschinica* Nekh. and with brachiopods *Anathyris ussovi* Khalf., *A. phalaena* Khalf., *Cyrtospirifer verneuili* (Murch.), *C. achmet* Nal., *Stropheodonta fischeri* (Vern.), *Schelwienella umbraculum* (Schloth.), *Productella subaculeata* (Murch.), *Lamellispirifer ales* Khalf., and *Mucrospirifer mesocostalis* (Hall.) (Rzhonsnitskaya, 1968, 1975; *Tipovye razrezy...*, 1992). According to characteristics of foraminifers and tabulate corals, the basin was shallow (Dubatolov, 1982; Poyarkov, 1969; Chuvashov, 1963). Hydrodynamics was active or moderately active. The water temperature was about 18–20°C.

The small Altai paleolandscape corresponded to the present-day areas of the Uimen'–Lebed' trough, Kosh-Agach and ore district of the Altai Mountains. In the southeast, it accumulated carbonate–terrigenous or terrigenous–carbonate deposits. The bottom relief was gently undulating, and the terrigenous deposits in the ore district were intercalated with acidic lava flows and tuffaceous beds. The active tectonics, intense hydrodynamics, and moderate water temperature characterized the shallow Altai basin, where the depth ranged from 10–20 to 50–60 m. The bottom illumination was sufficient for development of the diverse benthic faunas. The diversity maximum was in the northwestern area of the basin. The coral pavements at the bottom and biohermal to reefal buildups were typical of the Altai ore district.

The basin under consideration hosted the following groups of organisms: foraminifers *Nanicella* sp., *Bisphaera* cf. *elegans* Viss., *Tikhinella* sp., *Neoarchaesphaera* aff. *polypora* (Antrop), and *Suleimanovella suleimanovi* (Lip.) defined by Zadorozhnyi; stromatoporoids *Stachyodes* sp. and *Stromatoporella izylensis* Yavor. defined by V.G Khomykh; abundant and diverse tabulate corals represented by the early Frasnian (Gerikhovo) assemblage of *Pachyfavosites postremus* Dubat., *Thamnopora reticulata* (Blainv.), *Crassialveolites domrachevi* (Sok.), *C. grandis* Dubat., *Alveolites complanatus* Lec., and by the middle-late Frasnian ((Nikolaevka) assemblage of *Thamnopora cervicornis* (Blainv.), *Cladopora gracilis* Salee sensu Lec., *Trachypora* sp., *Coenites* sp., *Alveolites spasskyi* Dubat., *Crassialveolites domrachevi* (Sok.), and *Alveolitella* ex gr. *subaequalis* (M.-Edw. et H.). Chekalin and Zheltonogova (1985) described the more diverse coral assemblage from the Rubtsovsk region, where they identified, in addition to the listed forms, the tabulate species *Mesolites squamatus* (Dubat.), *Thamnopora boloniensis* (Goss.), *T. strelinaensis* Dubat., *patula* Tchud., *Gracilopora vermiculata* (McCoy), *Scoliopora denticulate* (M.-Edw. et H.), and *Mastopora compact* (Tchern.) associated with rugose corals *Hexagonaria arctic* (Meek.), *Macgeea multizonata* C. Reed, *M. pulchra* Spass., and *Tabulophyllum macconelli* Hail et Whit. Brachiopods from the same area represented genera *Gypidula, Euryspirifer, Mucrospirifer, Schizophoria, Spinatrypa, Desquamatia*, and *Adolfia*. Pelagic settings were inhabited by ammonoid genera *Ponticeras, Triainoceras, Manticoceras* and others (Bogoslovskii, 1958).

The early Frasnian (Gerikhovo) assemblage of rugose corals included *Neostringophyllum modicum* (Smith), *Macgeea solitaria* (M.-Edw. et H.), *M. multizonata* (Reed), *M. pulchra* Spass., *Megaphyllum longiseptatum* Ivania, and *Phillipsastraea thomasi* (Stein.). The middle and late Frasnian (Razdol'e and Nikolaevka) rugose assemblages consisted of *Nicholsoniella baschkirica* Soshk., *Tabulophyllum gorskii* Bulv., *Neostringophyllum modicum* (Smith), *Thamnophyllum virgatum* (Soshk.), *Macgeea multizonata* (Reed), and *Phillipsastraea filata* (Schloth.) (Spasskii, 1960).

Brachiopods were even more diverse than corals. Western areas of the Altai sea hosted 29 their species: 11 widespread in the early Frasnian substage and 20 in the middle and upper substages (Bublichenko, 1974). Most abundant among them were *Schizophoria septentrionalis* (Eichw.), *Gypidula cassidea* (Phill.), *Productella productoides* (Murch.), *Hypothyridina emmonsi* (Hall et Whit.), *Parapugnax altaicus* (Eichw.), *Spinatrypa tubaecostata* (Paeck.), *Desquamatia parazonata* (Kelus), *Adolfia laevigata* (Roem.), *Cyrtina triquetra* Hall (early Frasnian), *Nimphorhynchia* aff. *nympha* (Barr.), *Hypothyridina venustula* (Hall), and *Spinatrypa comitata* Copp. (middle–late Frasnian). The early Frasnian crinoid assemblage included *Trigonocyclicus* sp., *Hexacrinites kartsevae* Yelt. et Dubat., *H. argutus* (Yelt.), and *Anthinocrinus* sp. defined by Dubatolova and Zinevich.

In contrast, the late Frasnian crinoids were represented by *Lissocrinus* aff. *insectus* (Yelt.), *Schyschcatocrinus paucus* (J. Dubat.), *Callocrinus multicius* (J. Dubat.), and by five local species (determinations of Dubatolova).

Among trilobites, we should mention here *Scutellum alutaceum* Goldf., *S. longicandatum* Z. Max., *Acanthaloma (A.) tenella* Z. Max. (early Frasnian), *Scutellum* cf. *orientalis* Z. Max., *Phacops postalacius* Z. Max., *P.* aff. *pronini* Z. Max. (middle–late Frasnian according to Maksimova, 1960). In the ammonoid community, there were widespread 21 species (excluding those of the *Pharciceras* Horizon now dated back to the Givetian Age) of genera *Tornoceras, Ponticeras, Probeloceras, Manticoceras, Beloceras,* and *Devonopronorites*. Some of them, e.g., *Tornoceras simplex* (Buch), *Manticoceras carinatum* Beyr., *M. intumesens* (Beyr.), *M. sinuosum* (Hall), *M. galeatum* Wdkd., and *Beloceras sagittarium* (Sandb.) were also widespread in Frasnian seas of Eurasia, Africa, and North America. Conodonts represented another important group of fossils from the western part of the Altai sea (ore district). G.E. Shinkarev (personal communication) identified the following species among them: *Palmatolepis hassi* Mull. et Mull., *Pal. subrecta* Mull. et Young., *Pal. gigas* Mull. et Young., *Polygnathus decorosus* Stauff., *Pol. normalis* Mull. et Young., *Pol. aequalis* Klapp. et Lane, *Ancyrodella nodosa* Ulrich. et Bassl., *A. lobata* Brans,

et Mehl, *Icriodus symmetricus* Brans, et Mehl, and *Belodella devonica* (Stauff.). In the Rubtsovsk area, the conodont assemblage was even more diverse (Chekalin and Zheltonogova, 1985).

Rugose corals in the east of the Altai sea (Gornoaltaisk region) were less diverse than in the west, being represented by *Marisastrum sedgwicki* (M.-Edw. et H.), *M. carinata* (Bulv.), and by few other taxa. However, brachiopods were diverse here and included *Stropheodonta interstrialis* (Phill.), *Douvillina cayuta* Hall, *Plicochonetes nanus* (Vern.), *Productella subaculeata* (Murch.), *Mucrospirifer mesocostalis* (Hall), *Cyrtospirifer achmet* Nal., *C. schelonicus* Nal. var. *tenticuliformis* Ržon., *Atrypa waterlooensis* Webst., *A. uralica* Nal., *Retzia tschernyschewi* Peetz, *Athyris concentrica* Buch., *Anathyris phalaena* (Phill.), and *A. peetzi* Khalf. (*Biostratigrafiya paleozoya...*, 1960). *Fistulipora intermedia* Nekh., *Fenestella maxima* Krasn., *F. elongata* Krasn., *F. pioneri* Krasn., *Semicoscinium subquadratum* Krasn., *Reteporidra stellata* Krasn., *R. nataliae* Krasn., *R. carinata* Krasn., *Lioclema incrustans* Nekh., *L. polenovi* Nekh., *L. editum* Moroz., *L. vassinense* Moroz., *L. ramosa* Nekh., *Hemitrypa bugusunica* Nekh., *H. tubulosa* Nekh., *Saffordotaxis multispinata* Moroz, *Fistuliramus changi* (Yang.), *Semicoscinium angerlyki* Nekh., *S. delicatum* Krasn., and *S. subquadratum* Krasn. represented widespread bryozoan taxa (Volkova, 1974).

The Minusinsk lowland in the east was a source of miospores transported by winds into the Altai sea. The identified miospore species are *Archaeozonotriletes plicatus* Naum., *A. basilaris* Naum., *A. confusus* Naum. var. *medius* Tschibr., *Hymenozonotriletes limplickes* Naum., and *Lophozonotriletes crassatus* Naum. (*Resheniya...*, 1982). Genera *Izhella* and *Renalcis* represented the group of algae (data of V.A. Luchinina).

In general, the Altai sea of the Frasnian time was populated by very diverse benthic and pelagic organisms. It was well connected with sea basins of Eurasia and, probably, of North America, as one can judge from abundant remains of cosmopolitan taxa. Many of the organisms were thermophilic, and this, together with the character of sedimentation, suggests that the sea was situated inside the subtropical zone of that time.

In the Frasnian Age, the Taimyr paleolandscape extended consuming the Yamal–Gydan paleolandscape of the Givetian time. It accumulated calcareous, clayey–calcareous, and silty–calcareous ooze under quiet tectonic and hydrodynamic conditions. In general, sedimentation environments resembled those of the platform areas, because the fine horizontal lamination of sediments is almost intact, and their thickness is small. Clayey and calcareous ooze were deposited predominantly in the western paleolandscape areas (territory of the Yamal and Gydan peninsulas).

Fairly abundant benthic organisms were represented by the following groups of species: stromatoporoids *Amphipora ramosa* Phil.; tabulate corals *Thamnopora* sp., *Syringopora* sp., and *Pachyfavosites* sp.; brachiopods *Adolfispirifer jeremejevi* (Tschern.), *Chonetipustula petini* Nal., *Desquamatia tenuisulcata* Wen., *D. alicola* Frech, *Calvinaria taimyrica* Nal., *Cyrtospirifer* ex gr. *disjunctus* Sow., *Productella subaculeata* (Murch.), *Mucrospirifer novosibiricus* (Toll). Associated bivalves, bryozoans, and ostracods *Leperditia nalivkini* Lakk. were principal builders of coquinoid banks. Ammonoids *Manticoceras* sp. and *Tornoceras* sp. inhabited pelagic settings (*Resheniya...*, 1982; Bezzubtsev, 1989; Matukhin, 1991).

The depth range in the basin was from 10 to 100 m. The water mass of normal salinity was transparent, suitable for sufficient bottom illumination and fitting the life conditions of marine organisms.

The Central Siberian paleolandscape in the northwest of Siberian platform represented a shallow (few dozens of meters) sea that accumulated calcareous, clayey–calcareous, and dolomitic ooze together with sulfate deposits (Matukhin and Sokolov, 1991). In lithology of sediments, this paleolandscape was distinct from the West Siberian and Taimyr basins. Sedimentation developed here under conditions of the arid climate. Terrigenous sediments (clay, silt, and polymictic sand) were concentrated in the southwestern area of the paleolandscape. The greater part of the basin was populated by foraminifers, bryozoans, brachiopods, ostracods, and fishes. In contrast to other groups, foraminifers from the southwestern area are unknown. Remarkable species of foraminifers were *Moravammina* cf. *fragilis* E. Byk., *M. segmentata* Pok., *M. (Nodozinella)* aff. *tatarstanica* Antrop., *M. (Litya) sizranensis* E. Byk., *Evlania prava* Tchuv., *L. camerata* Tchuv., and *Vicinesphaera* cf. *squalida* Antrop. Among brachiopods, there were present *Chonetipustula* aff. *petini* Nal., *Productella* sp., *Spinatrypa* ex gr. *aspera* (Schloth.), *S. imangdini* Kryl., *Pugnax plicatus* Kryl., *P. menneri* Kryl., *Adolfispirifer jeremejewi* (Tschern.), and *Streptorhynchus devonicus* Orb. Species *Moelleritia crassa* Abusch., *M. tenue* Abusch., *Knoxiella* aff. *ichimensis* Pol., and *Bairdia fabaeformis* Pol. represented ostracods. The conodont assemblage included *Polygnathus decorosus* Stauf., *P. lanei* Kuzmin, *P. alatus* Huddle, and *P. webbi* Stauf. Concurrent vertebrate taxa were *Grossilepis* sp. and *Bothriolepis* sp. Species *Umbella bella* Masl., *U. bykovae* Reitl., and *U. baschkirica* E. Byk. exemplified charophytes. According to available data, the basin also hosted holothurians, bluegreen algae, and girvanells (Matukhin, 1991; Matukhin *et al.*, 1995).

This very diverse organic community characterizes the normal salinity of clear water in a shallow (10–20 m) basin, where substratum was soft. Sedimentation obviously progressed under quiet tectonic and hydrodynamic conditions.

The denudation lowlands of the Kazakhstan and Angarida continents were characterized by a slow erosion of siliceous slates. The terrigenous material from

these areas influenced sedimentation in those marine basins, where the clayey to clayey–calcareous material was an important component of deposits. Calcareous sandstones and siliceous phyllites were subjected to erosion in islands separating the Cis-Urals, East Uralian, and Kolyvan'–Tomsk paleolandscapes. The erosion was most intense in the western coast of the island bordering the Cis-Urals basin, where coarse terrigenous material (gravel and shingle) was deposited. On the other side of the island, this process was less active, and sedimentary sequences in the East Uralian paleolandscape include, in addition to calcareous mudstones, only silty and sandy beds also having the calcareous admixture.

The terrigenous material from the Ob' uplift, which included several islands, represented products of erosion of calcareous to siliceous slates and also of other crystalline rocks variable in composition and exposed within the gently undulating relief lacking contrast forms.

CONCLUSION

Summarizing data on marine paleolandscapes that existed in Siberia during the Eifelian, Givetian, and Frasnian ages, we arrived at the following conclusions:

(1) Like in the Early Devonian (Dubatolov and Krasnov, 1993, 1998), paleolandscapes of the Middle and Late Devonian time were differentiated with respect to sedimentation patterns and habitat environments of benthic organisms. In addition to the geographic position, these characteristic were controlled by the intensity of tectonic and magmatic activity in paleobasins.

(2) Diversity of benthic population in Siberian seas varied in space and with time. The most diverse species assemblages of the Middle Devonian–Frasnian period inhabited sea basins of the Altai–Sayan province. Within the West Siberian sea, the diversity maximum was typical of the Nyurol'ka–Var'egan and Kolyvan'–Tomsk paleolandscapes, whereas other areas of the sea were sparcely populated. As compared to the Eifelian benthic faunas, those of the Givetian Age became less diverse. First of all, this concerns stromatoporoids, tabulate corals, and bryozoans despite their abundance level comparable with the previous one. The diversity decrease continued in the Frasnian Age, and this trend is especially well evident in development of foraminifer assemblages. Algae widespread in many paleolandscapes reached their peak abundance at the Frasnian time in the Nyurol'ka–Var'egan, Central Siberian, and Kolyvan'–Tomsk paleolandscapes.

(3) Like in the Early Devonian, all subsequent sea basins of Siberia were situated within the subtropical arid zone. Proper conditions were especially distinct in the northwest of Siberian platform and Taimyr (Central Siberian and Taimyr seas), where sedimentary sequences include red beds and evaporite deposits. This conclusion well accords with distribution patterns of fruticose and arboreal plants (*Archaeopteris* and some other taxa) that were widespread, in addition to the indicated regions, also within a wide belt extending from the Elsmere Island, Alaska, and Oklahoma (United States) to the Spitsbergen Archipelago and further to China across the Altai–Sayan region and Kazakhstan (Snigirevskaya, 1988; Nadler and Kuznetsova, 1995). Typical of the entire belt was accumulation of red lacustrine and alluvial deposits along with deposition of gypsum beds especially widespread at the Givetian level. Many biohermal and reefal massifs were concentrated within this belt, which extended parallel to the western Urals. Another subtropical belt with abundant coral reefs was located westward spanning Europe and Appalachian Mountains of North America (Pajchlowa and Stasinska, 1965; Lecompte, 1970; Tsien, 1974, 1980; Oliver and Pedder, 1979; Burchette, 1981).

(4) Variable tectonic and volcanic activity was an important feature of Siberian seas. As before, the tectonic activity of the Eifelian time was concentrated in western areas of the West Siberian sea. This is evident from frequent lava eruptions within the narrow paleolandscape zones oriented along the meridional direction in this part of the sea. Volcanic eruptions of acidic lavas are also recorded in the Salair region and in the south of the West Siberian sea. The Givetian Age was a time of volcanism cessation in the western paleolandscapes, but the volcanic activity was still in progress inside the Salym basin. Recommencement of intense volcanic eruptions at the Frasnian time was characteristic of western and central areas of the West Siberian sea, and also of the Yamal–Gydan and Salair–Minusinsk paleolandscapes; within the latter, they took place in the western part.

(5) The shallow sea basin and their biotas inherited from the Early Devonian time became interrelated better and had improved connections with other neighboring and remote basins, the Chinese basins and those of Indochina included (Liou, 1962). Most diverse benthic assemblages populated the Nyurol'ka–Var'egan, Kolyvan'–Tomsk, and other paleobasins located at the sites of the present-day Salair, Kuzbass, and Altai regions.

(6) The late Givetian transgression spread over the vast areas of Siberia and exerted a strong influence on biota development and sedimentation in Siberian basins. Its impact was most significant in the Salair–Minusinsk sea that was situated in the southeast of the Altai–Sayan Mountains (Krasnov, 1970), in the northwestern areas of the Siberian platform, and also in the Taimyr region. The differentiated predominance of various faunal groups in Siberian seas suggests that seawaters advanced in particular basins of Siberia from different sides, e.g., from Mongolia and China into the Salair–Minusinsk sea, from Europe into the western areas of the West Siberian sea, and from northeastern Russia into the Central Siberian and Taimyr seas.

(7) The general deepening of sea basins in response to the late Givetian transgression was less pronounced in central areas of the West Siberian sea, but accumulation of saliferous deposits considerably decreased even in the Central Siberian basin.

(8) Coastal areas of lacustrine–alluvial plains, which bordered Siberian seas in the east, accumulated red and variegated terrigenous deposits. These areas were favorable for development of vegetation (Anan'ev, 1959).

(9) During the entire Middle Devonian, land masses of Kazakhstan, Angarida, and several islands were important elements of paleolandscape zoning in Siberia. Like in the Emsian Age, the Kazakhstan continent of the Eifelian time comprised the hilly uplift in the east and the denudation plain at its western and northern margins. At this time, its headland jutted northward into the West Siberian sea reaching the latitude of the Khanty-Mansiisk. Later, during the Givetian Age, the whole area of the continent represented the alluvial denudation plain; it turned again into the hilly uplift during the Frasnian regression. The Angarida land of the Eifelian time was mountainous in the south, corresponding to the denudation plain in central and western areas. The late Givetian transgression and concurrent crustal subsidence turned the entire land area into the denudation plain. In the Frasnian age, the Angarida area represented again the hilly land. Its coastal zone was covered with fruticose–arboreal vegetation and bordered the Central Siberian sea in the north and northwest.

(10) Biohermal and reefal buildups of Siberian seas extended as a belt from the Salair–Minusinsk region across the Nyurol'ka–Var'egan basin into the Kolyvan'–Tomsk sea. The buildups were small, characteristic of shallow sea basins. Stromatoporoids and algae together with tabulate and rugose corals were main groups of reefal frame-builders (Krasnov *et al.*, 1986; Stepanov *et al.*, 1990). Algae became leaders in this respect during the Late Devonian.

Reviewers M.A. Zharkov and B.S. Sokolov

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Triassic Radiolarian Zonation in the Far East of Russia

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Abstract—In siliceous deposits of eastern Russia we can recognize, by means of radiolarians, the following biostratigraphic units: *Pseudostylosphaera fragilis* Beds (upper Olenekian) and zones of *Hosmadia gifuensis* (lower Anisian), *Triassocampe diordinis* (middle Anisian), *Triassocampe deweveri* (upper Anisian), *Triassocampe scalaris* (uppermost Anisian–lowermost Ladinian), *Oertlispongus inaequispinosus* (the lower part of the lower Ladinian), *Falcispongus falciformis* (the upper part of the lower Ladinian–lower part of the upper Ladinian), *Muelleritortis cochleata* (the middle–upper part of the upper Ladinian), *Tritortis kretaensis kretaensis* (lower Carnian), *Capnuchosphaera theloides* (upper Carnian), *Capnodoce crystallina* (lower–middle Norian), *Lysemelas olbia* (the lower part of the upper Norian), *Betraccium deweveri* (the upper part of the upper Norian), *Livarella densipora* (lower Rhaetian), and *Globolaxtorum tozeri* (upper Rhaetian). All zones distinguished in sections of siliceous monofacies are well correlative to those in carbonate and terrigenous sections far away. The detailed and persistent radiolarian biozonation of the Middle–Upper Triassic suggest a possibility to use this paleontological group not only for concrete stratigraphic purposes (e.g., for dating siliceous sequences), but also in a wider aspects of the global and regional Triassic stratigraphy.

Key words: **radiolarians, Triassic, zones, correlation, facies, Pacific belt, Mediterranean belt.**

The use of radiolarians for the Triassic zonation began 20 years ago. In the late 1970s and early 1980s, the initial regional radiolarian-based zonal schemes were elaborated for the Upper Triassic siliceous deposits of Mexico (Pessagno *et al.*, 1979) and the Middle–Upper Triassic of Japan (Yao, 1982). Resolution of these schemes was not high, as zones were equivalent in range to stages or larger intervals. The zones were distinguished by means of typical radiolarian assemblages, but some adjacent zones (e.g., the *Triassocampe deweveri* and *T. nova* zones of Japan) were lacking common species (Yao, 1982), and this was the reason of subsequent doubts about the successive character of relevant assemblages (Bragin, 1986).

The first half of the 1980s was marked by elaboration of detailed Triassic radiolarian zonations for both siliceous and terrigenous deposits. For instance, the Upper Triassic terrigenous deposits of Oregon and British Columbia were subdivided into two zones and five subzones (Blome, 1984). Detailed radiolarian zonations were elaborated also for the Middle–Upper Triassic siliceous deposits of Japan (Sato, 1986; Yoshida, 1986) and Sakhalin (Bragin, 1986). New materials obtained from earlier unstudied stratigraphic intervals (upper Ladinian, lower Carnian) provided better knowledge of radiolarian stratigraphy and grounded a more reliable subdivision and correlation of host deposits. However, there was still a problem of dating the newly found radiolarian assemblages, because they were often described from siliceous sections, stratigraphy of which was insufficiently studied.

In the late 1980s and the first half of the 1990s, the problem was solved in general owing to thorough investigations of radiolarians from ammonoid-bearing carbonate and terrigenous sections (Carter, 1993; Kozur and Mostler, 1994) and due to the refined subdivision of Triassic siliceous deposits on the basis of conodonts with consequent correlation of conodont- and radiolarian-based stratigraphic units (Bragin, 1991). This allowed the existing schemes to be modified and correlated to each other (Bragin, 1991; Sugiyama, 1992, 1997; Carter, 1993; Kozur and Mostler, 1994; Blome *et al.*, 1995; Kozur *et al.*, 1996a, 1996b) (table).

The current stage of studies on the Triassic radiolarian biostratigraphy is as follows. Several zonations are elaborated for the Triassic System, but many of them do not characterize the system entirely, and some correspond to a series or even a stage only (Blome, 1984; Carter, 1993). In addition, the zonal scales published are unreliably correlated to each other and to standard subdivisions. Finally, paleontological characteristics of almost all radiolarian zonations should be verified, because numerous Triassic radiolarian taxa have been described recently and data on the radiolarian phylogeny and evolution are insufficiently used. All this determines principal lines of current investigations. In general, the main goal is to develop the Triassic biostratigraphy on the basis of comprehensively studied radiolarian evolution during the Triassic.

An initial picture of evolutionary stages of Triassic radiolarians was outlined more than 10 years ago (Bragin, 1988). In short, their evolution was subdivided into three major stages: Early Triassic (the time of extremely low taxonomic diversity of radiolarians in response to their great extinction at the end of the Permian), Middle Triassic (the time when conservative Paleozoic taxa

Triassic radiolarian zonations

Stage	Substage	Bragin (this article), Southern Sikhote-Alin'	Sigiyma, 1997, Japan	Kozur et Mostler, 1994, West Europe
Rhaetian	Upper	*Globolaxtorum tozeri*	*Haeckelicyrtium breviora*	
	Lower	*Livarella densiporata*	"Skirt F"	*Livarella densiporata*
Norian	Upper	*Betraccium deweveri*	*Praemesosaturnalis pseudokahleri*	
			Praemesosaturnalis multidentatus	
		Lysemelas olbia	*Lysemelas olbia*	
	Middle	*Capnodoce crystallina*	*Trialatus robustus–L. olbia*	*Capnodoce ruesti*
	Lower		*Capnodoce–Trialatus*	
Carnian	Upper	*Capnuchosphaera theloides*	*Poulpus carcharus*	*Nakasekoellus inkensis*
			Capnuchosphaera	*Tetraporobrachia haeckeli*
	Lower	*Tritortis kretaensis kretaensis*	*Spongoserrula dehli*	*Tritortis kretaensis*
Ladinian	Upper	*Muelleritortis cochleata*	*Muelleritortis cochleata*	*Muelleritortis cochleata*
	Lower	*Falcispongus falciformis*	*Yeharaia elegans*	*Ladionocampe multiperforata*
		Oertlispongus inaequispongus		*Spongosilicarmiger italicus* / *O. inaequispinosus*, *O. primitivus*, *S. i. transitus*
		Triassocampe scalaris		
Anisian	Upper	*Triassocampe deweveri*	"Spine A2"	
			Triassocampe deweveri	*Tetraspinocyrtis laevis*
	Middle	*Triassocampe coronata*	*Triassocampe coronata*	*Parasepsagon robustus*
	Lower	*Hozmadia gifuensis*	*Eptingium nakasekoi*	
Olenekian	Upper	*Pseudostylosphaera fragilis*	*Parentactinia nakatsugawaensis*	
			Follicucullus	

coexisted with newly appeared progressive ones), and Late Triassic (the epoch of new taxa predominance and formation of typical Mesozoic fauna). Afterward, the correctness of this idea was proved, and new possibilities to detail the radiolarian-based stratigraphy, first of all, for the Middle–Upper Triassic appeared.

Taxa used most frequently to recognize biostratigraphic units are those having their first occurrences in the Triassic and distinguished from earlier forms by the essentially different morphology (Bragin, 1991, 1999). These are families Triassocampidae, Oertlispongiidae, Muelleritortiidae, Capnuchosphaeridae, Pantanelliidae, and Saturnalidae. Most of index-species and characteristic taxa of biostratigraphic units recognized here belong to these families. The distinguished zones are in correspondence with evolutionary stages of these groups and sometimes with morphogenesis of individual taxa.

A good example is the family Oertlispongiidae characteristic of the upper Lower Triassic, the entire Middle Triassic, and the lower Carnian. The family

demonstrates changes in morphology of main spines from smooth and straight at the beginning to curved, then to blade-shaped, and finally, to lobate (Bragin, 1999). Due to a rapid substitution of allied morphotypes, several biostratigraphic units can be recognized in the Middle Triassic (table). Another example is the genus *Triassocampe*, which shows the following phylomorphogenetic succession: *Triassocampe coronata—T. deweveri—T. scalaris—T. sulovensis* (Bragin, 1999) also taken into account in the Triassic zonation in the Far East of Russia (table).

Analyzing the Triassic radiolarian evolution, it is possible to establish many well-traceable levels and intervals, where stratigraphically important taxa had their first or last occurrences (Bragin, 1999):

(1) The uppermost Olenekian–lowermost Anisian or the appearance period of the genus *Hozmadia* (monocyrtoid nassellarians); the interval can be recognized in the Dal'negorsk and Khabarovsk areas of Sikhote-Alin' (Bragin, 1991, 1992) and in the Kinkazan Mountain sections of central Japan (Sugiyama, 1992).

(2) The lower–middle Anisian boundary corresponding to the appearance level of the genus *Pararuesticyrtium* (the earliest multicyrtoid nassellarians) and the nassellarian families Tripedurnulidae and Eptingiidae; the level is defined in the Dal'negorsk area of Sikhote-Alin' (Bragin, 1991).

(3) The middle part of the middle Anisian, when the earliest *Triassocampe* forms (*Triassocampe coronata* and *T. diordinis*) appeared; the interval is recognizable in the Dal'negorsk and Khabarovsk areas of Sikhote-Alin' (Bragin, 1991, 1992) and in the Kinkazan and Inuyama sections of Japan (Sugiyama, 1992; Matsuoka *et al.*, 1994).

(4) The middle–upper Anisian boundary marking the appearance of advanced *Triassocampe* forms (*T. deweveri*), *Hindeosphaera* genus, and families Relindellidae and Pentactinocarpidae; the level is detected in the Dal'negorsk and Khabarovsk areas of Sikhote-Alin' (Bragin, 1991, 1992), in the Kinkazan and Inuyama sections of Japan (Sugiyama, 1992; Matsuoka *et al.*, 1994), and in the Alpine–Mediterranean region (Kozur and Mostler, 1994).

(5) The upper Anisian–lower Ladinian interval, when higher *Triassocampe* forms (*T. scalaris*) and progressive Oertlispongiidae (*Oertlispongus inaequispinosus*) appear in association with genera *Pylostephanidium, Yeharaia, Silicarmiger*, and *Triassobipedis*; the interval is recognized in the Dal'negorsk area of Sikhote-Alin' (Bragin, 1991), Inuyama area of Japan (Yao, 1982; Matsuoka *et al.*, 1994), the Koryak Upland (Bragin, 1991), and northern Italy (Kozur and Mostler, 1994).

(6) The middle part of the lower Ladinian or the appearance time of Oertlispongiidae forms with flat spines (genera *Falcispongus* and *Baumgartneria*); the level is recognized in the Dal'negorsk area of Sikhote-Alin' and northern Italy (Kozur and Mostler, 1994).

(7) The middle part of the upper Ladinian characterizing the appearance interval of Muelleritortiidae, higher Oertlispongiidae forms (genera *Spongoserrula* and *Pterospongus*), and Austrisaturnalidae with parallel extinction of *Triassocampe deweveri* species; the interval is distinguished in the Dal'negorsk area of Sikhote-Alin' (Bragin, 1991), southern Sakhalin (Bragin, 1986, 1991), Hungary and Austria (Kozur and Mostler, 1994).

(8) The Ladinian–Carnian boundary corresponding to the initial dominance of the genus *Tritortis* (Muelleritortiidae) and to appearance level of the genera *Relindella* and *Pentaspongodiscus*, and also of the early *Xiphotheca* forms; the level is recognized in the Dal'negorsk area of Sikhote-Alin' (Bragin, 1991), southern Sakhalin (Bragin, 1986, 1991), Hungary and Austria (Kozur and Mostler, 1994).

(9) The upper part of the lower Carnian, when of the earliest Saturnalidae (*Palaeosaturnalis*) appeared, but Oertlispongiidae and Muelleritortiidae became extinct; the interval is recognized in Hungary (Kozur and Mostler, 1994), southern Sakhalin, and Turkey.

(10) The middle part of the upper Carnian, when families Capnuchosphaeridae and Capnodocidae appeared along with nassellarian genera *Japonocampe, Multimonilis*, and *Latium*; the interval is recognized in the Dal'negorsk area of Sikhote-Alin' (Bragin, 1991), southern Sakhalin (Bragin, 1986, 1991), and Japan (Yao, 1982; Matsuoka *et al.*, 1994).

(11) The Carnian–Norian boundary marking the initial bloom of subfamily Capnodocinae, extinction of Austrisaturnalidae, and appearance of genera *Squinabolella* and *Nabolella*; the level is recognized in the Dal'negorsk area of Sikhote-Alin' (Bragin, 1991), southern Sakhalin (Bragin, 1986, 1991), and the Koryak Upland (Bragin, 1991, 1992).

(12) The middle–upper Norian boundary or the level marking the first occurrence of subfamily Pantanelliinae and genera *Livarella, Praemesosaturnalis*, and *Pseudoheliodiscus*, simultaneous extinction of family Capnuchosphaeridae, subfamily Capnodocinae, and genus *Triassocampe*, as well as the prosperity time of Ferresiidae; the level is recognized in the Dal'negorsk area of Sikhote-Alin' (Bragin, 1991), Southern Sakhalin (Bragin, 1986, 1991), the Koryak Upland (Bragin, 1991a, 1992), Japan (Yao, 1982; Yoshida, 1986; Matsuoka *et al.*, 1994), Oregon (Blome, 1984), and British Columbia (Blome, 1984; Carter, 1993).

(13) The Triassic–Jurassic boundary corresponding to extinction level of genera *Livarella, Pentactinocarpus, Ferresium, Risella*, and *Eptingium* (Bragin, 1994); the level is recognized in Japan (Yao, 1982; Yoshida, 1986; Matsuoka *et al.*, 1994), Oregon (Blome, 1984), and British Columbia (Blome, 1984; Carter, 1993).

These reliable marker intervals and levels show that the suggested zonal scheme is well substantiated. However, the distinguished biostratigraphic units are well recognizable only in the warm-water paleogeographic

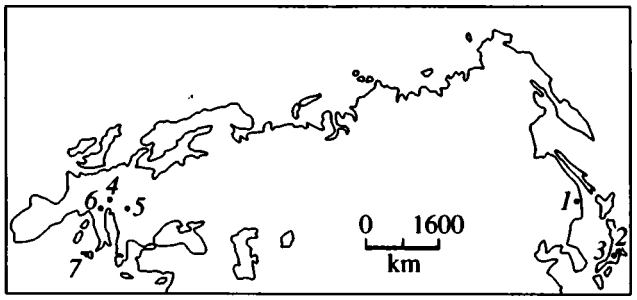

Fig. 1. The Triassic reference sections of Eurasia used as basis for radiolarian zonations: (*1*) Dal'negorsk of Sikhote-Alin'; (*2*) Kinkazan of Japan; (*3*) Inuyama of Japan; (*4*) Zlambachgraben of Austria; (*5*) Koveskal and Dallapuzhta of Hungary, Balaton Highland; (*6*) Passo della Gabiola of Italy; (*7*) Monte Cammarata of Italy, Sicily.

regions. The Boreal and Notal radiolarian assemblages of the Triassic from the Omolon massif and New Zealand have essentially different taxonomic composition (Aita, 1994; Bragin, 1994; Egorov and Bragin, 1995). The radiolarian-based correlation of boreal and tropical sections is a subject of future investigations.

This article presents the Triassic radiolarian biostratigraphic units recognized in siliceous sections, which have their own merits and shortcomings. The undoubted advantage over carbonate and terrigenous deposits is that they are of the monofacies type being characterized throughout by highly diverse radiolarian assemblages. Accordingly, we may establish biostratigraphic succession within a considerable stratigraphic interval showing only minor facies changes. On the contrary, in terrigenous and carbonate sections accumulated in shallower settings, radiolarians are often confined to separate short intervals or even to local levels, and changes in their assemblages are frequently caused by the environmental factors (e.g., by transition from nearly oceanic to neritic conditions and *vice versa*).

The main shortcoming of siliceous sections is their complicated geological structure. Mesozoic siliceous sections in mobile belts experienced different tectonic dislocations difficult for recognition and responsible for origin of frequent gaps or repeating intervals. The difficulty can be surmounted by a thorough investigation and correlation of undisturbed intervals in the sections in order to reconstruct their original succession. This approach was proved to be correct by means of independent and well compatible data on siliceous, carbonate, and terrigenous facies (Bragin, 1991; Carter, 1993; Kozur and Mostler, 1994; Blome *et al.*, 1995; Sugiyama, 1997).

The suggested zonation represents a considerably modified scheme for the Triassic siliceous deposits of eastern Russia as compared to that published more than 10 years ago (Bragin, 1988) (table). As before, the Triassic siliceous section of the Dal'negorsk area in Sikhote-Alin' (Bragin *et al.*, 1988; Bragin, 1991) was taken as the reference one (Figs. 1–3). The current investigations of the Dal'negorsk section confirm in general the previous conclusions, though the modifications increase the resolution of the scheme. The Dal'negorsk section is correlated to other siliceous sections in the Pacific belt, including Japan, and to carbonate and terrigenous sections located occasionally very far away (table, Fig. 1).

THE TRIASSIC RADIOLARIAN ZONATION

OLENEKIAN STAGE

Pseudostylosphaera fragilis Beds

Index-species: Pseudostylosphaera fragilis (Bragin) (="*Stylosphaera*" *fragilis* Bragin, 1991, p. 89, Plate 1, Fig. 4).

Type area: Sikhote-Alin', the Triassic section at the Dal'negorsk Settlement, Beds 4, 5 (Bragin *et al.*, 1988; Bragin, 1991, with modifications).

Lithology: (Figs. 2, 3):[1] Red, crimson- and brick-red indistinctly bedded, locally indistinctly flaggy siliceous mudstones with interbeds of light gray–green siliceous mudstones intercalated with frequent thin interbeds of light bluish gray and light greenish gray translucent glassy cherts, and with rare interbeds of black carbonaceous mudstones (at the top of the unit). Thickness is 8.5 m.

Nomenclature. Stratigraphic range of the unit is as previously established (the upper Olenekian).

Stratigraphy and paleontological characteristics. The lower boundary of the beds is drawn at the first occurrence level of index-species in association with *Hozmadia ozawai* Sugiyama, *Parentactinia nakataugawaensis* Sugiyama, and *Pseudostylosphaera kozuri* Sugiyama (Fig. 3). Slightly above there appear *Cryptostephanidium japonicum* (Nakaseko et Nishimura), *Parasepsagon* sp. cf. *P. longidentatum* (Kozur et Mostler), *Spongostephanidium longispinosum* Sashida, and *Eptingium nakasekoi* Kozur et Mostler. The beds also yield bilaterally symmetrical forms of the genus *Follicucullus* and poorly studied abundant forms with large rod-shaped spines, which may represent the early Paurinellidae (spumellarians).

Age substantiation. Beds 4 and 5 of the Dal'negorsk section are characterized by conodonts typical of the *Neospathodus triangularis* and *N. homeri* zones of the upper Olenekian, or, according to the other terminology, of the Spathian Stage (Bragin, 1991). The siliceous sections with similar paleontological characteristics are known near the Sadovyi Settlement and Khabarovsk (Bragin, 1991, 1992), and also in the Inuyama area of central Japan (Sugiyama, 1992; Matsuoka *et al.*, 1994; Kamata, 1995).

Note. Being well traceable, this unit cannot be ranked as a zone because the underlying beds are

[1] Lithology is described only for biostratigraphic units distinguished in the Dal'negorsk stratotype.

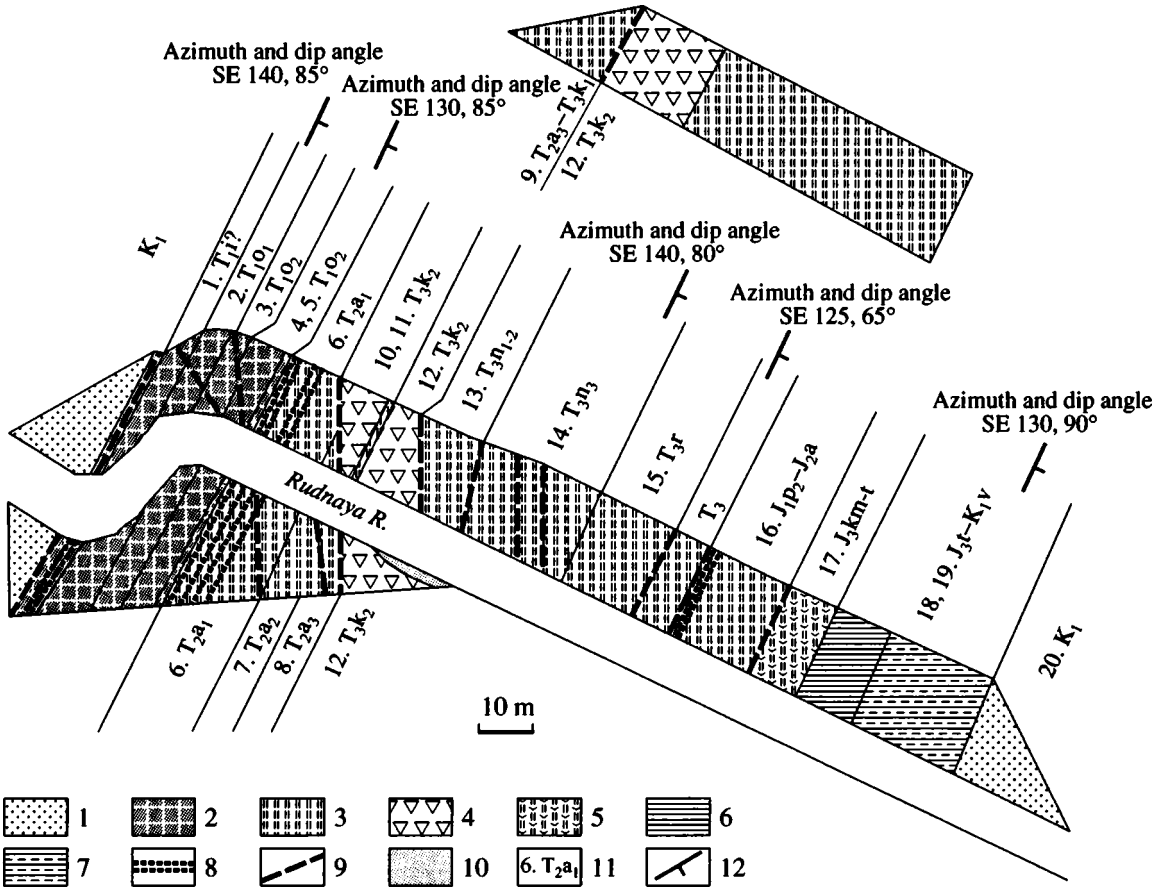

Fig. 2. The Dal'negorsk section of siliceous, volcanogenic and terrigenous Triassic–Lower Cretaceous deposits: (1) sandstone; (2) siliceous mudstone; (3) chert; (4) basalt, diabase, and hyaloclastite; (5) siliceous tuffite; (6) mudstone with siltstone interbeds; (7) siltstone with mudstone and sandstone interbeds; (8) horizons of carbonaceous rocks; (9), fault; (10), dike of intermediate composition; (11) bed number and index; (12) bed attitude.

devoid of representative radiolarians. Radiolarians of the Lower Triassic, especially the Induan and the lower Olenekian, have extremely poor taxonomic composition and are known from few localities. As many other organic groups, radiolarians became almost extinct at the end of the Permian and gained the taxonomic diversity very slowly during the Early Triassic time (Bragin, 1991; Kozur *et al.*, 1996b).

Distribution. Sikhote-Alin' is the type region (Bragin, 1991). In Japan this stratigraphic interval includes the *Follicucullus* and *Parentactinia nakatsugawaensis* zones (Sugiyama, 1992). The radiolarian assemblage of the *Parentactinia nakatsugawaensis* Zone is here similar in composition to that of the *Pseudostylosphaera fragilis* Beds but more diverse owing to occurrence of the genera *Poulpus* and *Nofrema* (nassellarians). Assemblages of the *P. nakatsugawaensis* Zone are widespread in central and southwestern Japan (Sashida, 1983, 1991; Sugiyama, 1992; Kamata, 1995).

LOWER ANISIAN

Hozmadia gifuensis Zone

Index-species: Hozmadia gifuensis Sugiyama (Sugiyama, 1992, p. 1194, Figs. 9, 6–8).

Type area: central Japan, Gifu Prefecture, the Kinkazan Mountain (Sugiyama, 1992).

Nomenclature. The *Hozmadia* Genus-zone of the same scope was previously established in Sikhote-Alin' (Bragin, 1991). After describing a number of new species of this genus, Sugiyama (1992) recognized the *Hozmadia gifuensis* Zone virtually correspondent to that mentioned above. It should be noted that the lower zonal boundary appears coincident in Japan with the Lower–Middle Triassic boundary, because the lower Anisian age of the conodont *Neospathodus timorensis* Zone is widely accepted (Sweet and Bergstrom, 1986; Bragin, 1991; Sugiyama, 1992; Matsuoka *et al.*, 1994).

Stratigraphy and paleontological characteristics. The lower boundary is drawn at the first occurrence level of the index-species and associated *Celluronta donax* Sugiyama (Fig. 3). Many species, such as *Eptingium nakasekoi* Kozur et Mostler, *Cryptostephanidium*

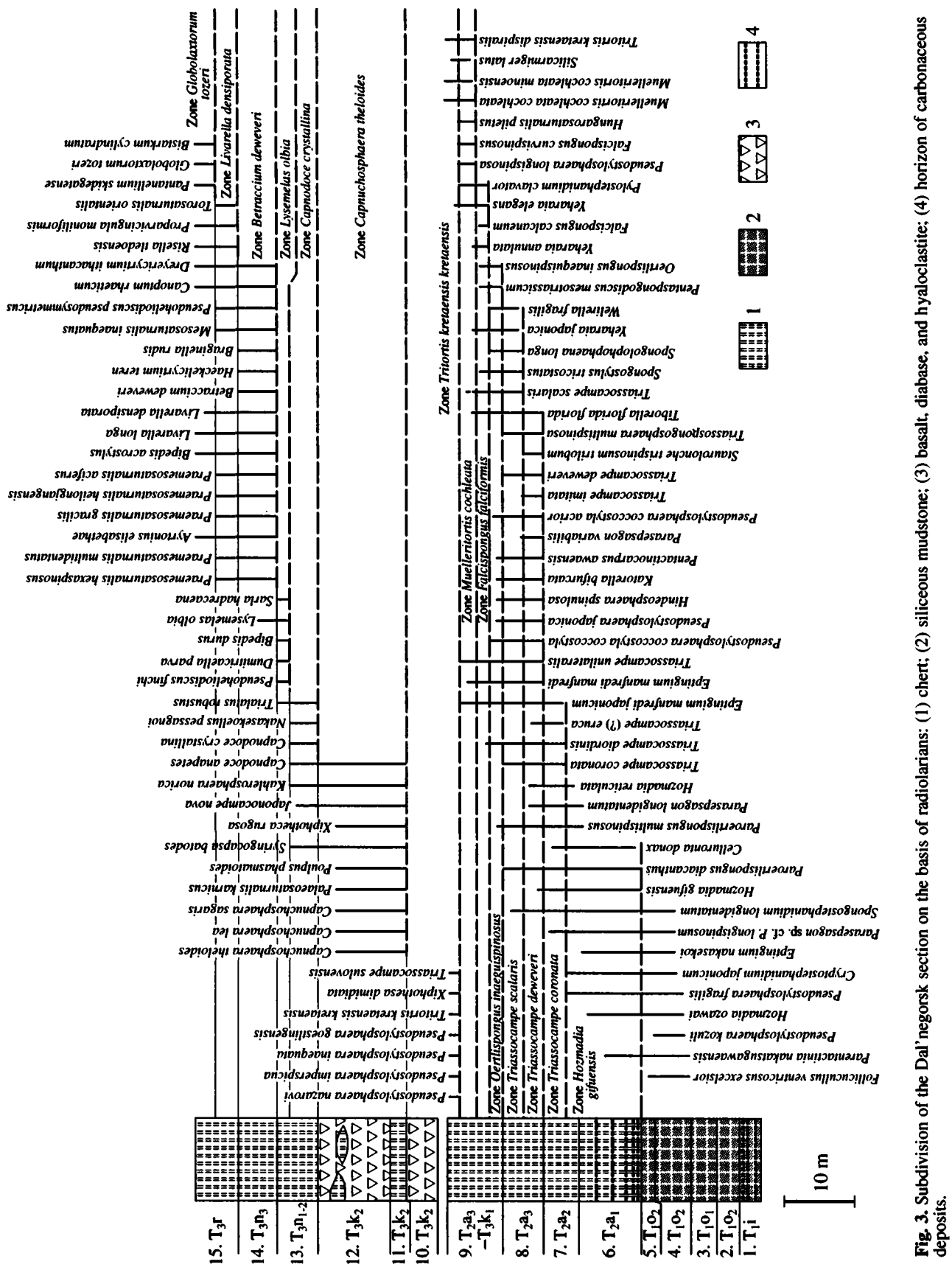

Fig. 3. Subdivision of the Dal'negorsk section on the basis of radiolarians: (1) chert; (2) siliceous mudstone; (3) basalt, diabase, and hyaloclastite; (4) horizon of carbonaceous deposits.

japonicum, Parentactinia nakatsugawaensis, Paroertlispongus diacanthus, Pseudostylosphaera fragilis, and *Spongostephanidium longispinosum,* are inherited from the underlying beds. In the Dal'negorsk section (Bragin *et al.*, 1988), the upper part of the zone contains *Paroertlispongus multispinosus* Kozur et Mostler, *Hozmadia rotunda* (Nakaseko et Nishimura), and *H. reticulata dumitrica* Kozur et Mostler (Bragin, 1991). *Tiborella* sp. and *Triassocampe* (?) sp. also have their first occurrences in this zone of the Dal'negorsk section.

Age substantiation. In the Dal'negorsk and Khabarovsk sections, this interval encloses conodonts of the *Neospathodus timorensis* Zone (Bragin, 1991) universally attributed to the uppermost Olenekian–lower Anisian (Sweet, 1970; Sweet *et al.*, 1971; Clark *et al.*, 1979; Kozur, 1980; Koike, 1981; Sashida and Igo, 1992). In the Kinkazan section (Sugiyama, 1992), the corresponding deposits yielded no conodonts, and the zone scope was determined according to its position between reliably dated deposits of the lower Triassic and the middle Anisian.

Note. In Sikhote-Alin' and Japan, the early Anisian time was marked by the increasing silica accumulation and rapid diversification of radiolarians. Nassellarians were previously considered to appear at that time (Bragin, 1991), but now they are proved to occur in the upper beds of the Lower Triassic (Sugiyama, 1992; Kamata, 1995).

Distribution: central Japan (Sugiyama, 1992; Matsuoka *et al.*, 1994) and Sikhote-Alin' (Bragin, 1991a, 1992); similar radiolarian assemblage was found in Thailand (Sashida and Igo, 1992).

MIDDLE ANISIAN

Triassocampe coronata Zone

Index-species: Triassocampe coronata Bragin (Bragin, 1991, p. 99, Plate 1, Fig. 15).

Type area: central Japan, Gifu Prefecture, the Kinkazan Mountain (Sugiyama, 1992).

Nomenclature. An analogous *Triassocampe diordinis* Zone embracing the entire middle Anisian was previously established in Sikhote-Alin' (Bragin, 1991). Later the *Triassocampe coronata* Zone of the same scope was recognized in Japan (Sugiyama, 1992). The well-identifiable species *T. coronata* appears to be more suitable for the zonal index-species. The lower boundary cannot be precisely defined in the type area because of abnormal stratigraphic contact between this zone and underlying beds (Sugiyama, 1992). In the Dal'negorsk section, the *T. coronata* Zone rests immediately on the lower Anisian *Hozmadia gifuensis* Zone.

Stratigraphy and paleontological characteristics. The lower boundary of the zone is drawn at the first occurrence level of the index-species associated with *Eptingium manfredi japonicum* (Nakaseko et Nishimura), *Hindeosphaera spinulosa* (Nakaseko et Nishimura), *Pararuesticyrtium imitatum* (Bragin), *Pseudostylosphaera compacta* (Nakaseko et Nishimura), *Triassocampe* sp. aff. *T. diordinis* Bragin, *T. diordinis* Bragin, and *T. (?) eruca* Sugiyama (Fig. 3). The zone contains *Hozmadia gifuensis, H. rotunda, Paroertlispongus diacanthus, Parasepsagon antiquum* (Sugiyama), and *Katorella bifurcata* Kozur et Mostler, all inherited from the preceding zone. Species *Celluronta donax* has its last occurrence in the upper part of the zone.

Age substantiation. In the sections of Sikhote Alin' (Dal'negorsk, Vysokogorsk) and Japan (Inuyama, Kinkazan Mountain, Shikoku Island), this interval yields the middle Anisian conodont species *Neogondolella bulgarica.* In the Dal'negorsk section, this species appears before the characteristic assemblage of the *Triassocampe coronata* Zone. For this reason, the lower boundary of the zone in the suggested scheme does not coincide with the lower–middle Anisian boundary, as it does in Japan (Sugiyama, 1997). It lies within the middle Anisian (Fig. 3, table).

Note. In the Pacific belt sections, the stratigraphic interval in question is characterized by the first occurrence of abundant and diverse forms of the genus *Triassocampe* (multicyrtoid nassellarians). These primitive forms have poorly sculptured test segments and one or two pore sets. In the Tethyan area, representative of *Triassocampe* occur sporadically probably because of relative isolation of the Tethys from the Paleopacific basins. The Mediterranean *Parasepsagon robustus* Zone spanning the same interval (Kozur and Mostler, 1994) is characterized by radiolarian assemblage essentially different in taxonomic composition from that of the *Triassocampe coronata* Zone in Sikhote-Alin' and Japan. It is still difficult to correlate these units.

Distribution: central and southwestern areas of Japan (Ishida, 1984; Sugiyama, 1992; Matsuoka *et al.*, 1994), Sikhote-Alin' (Bragin, 1991).

UPPER ANISIAN

Triassocampe deweveri Zone

Index-species: Triassocampe deweveri (Nakaseko et Nishimura) (=*Diotyomitrella deweveri* Nakaseko et Nishimura, 1979, p. 77, Plate X, Figs. 8, 9).

Type area: central Japan, Gifu Prefecture, the Inuyama section (Yao, 1982).

Nomenclature. The zone was originally established in the scope of the Ladinian Stage (Yao, 1982). Later it was shown (Bragin, 1991) that the index-species and some accompanying forms appear for the first time in the upper Anisian deposits, whereas the upper Ladinian beds are characterized by radically different radiolarian assemblages. Accordingly, the zone was considered as corresponding to the interval of the upper Anisian–lower Ladinian (Bragin, 1991). Further, the upper part of the zone was recognized as an independent *Triasso-*

campe scalaris Zone. The initial scope of the *T. deweveri* Zone was reduced to the lower part of the upper Anisian.

Stratigraphy and paleontological characteristics. The lower boundary is drawn at the first occurrence level of the index-species and associated *Eptingium manfredi manfredi* Dumitrica, Kozur et Mostler, *Parasepsagon variabilis* (Nakaseko et Nishimura), *Pseudostylosphaera japonica* (Nakaseko et Nishimura), *P. coccostyla acrior* (Bragin), *P. coccostyla coccostyla* (Rust), *Pentactinocarpus fusiformis* Dumitrica, *P. awaensis* (Nakaseko et Nishimura), *Staurolonche trispinosum trilobum* (Nakaseko et Nishimura), *Triassistephanidium anisicum* Kozur, Krainer et Mostler, and *Triassospongosphaera multispinosa* Kozur et Mostler (Fig. 3). *Triassocampe coronata, T. diordinis, Hozmadia rotunda, Eptingium manfredi japonicum*, and *Hindeosphaera spinulosa* still occur in the zone, but *Hozmadia gifuensis* disappears.

Age substantiation. In the Sikhote-Alin' sections, this stratigraphic interval contains conodonts of the upper Anisian *Neogondolella excelsa* Zone (Bragin, 1991). Mediterranean analogues of the zone enclose the late Anisian ammonoids and conodonts (Kozur *et al.*, 1996a). In the scheme suggested, the lower boundary of the *T. deweveri* Zone does not lie within the middle Anisian deposits, as it does in Japan (Sugiyama, 1997), but coincides with the middle–upper Anisian boundary (Fig. 3, table). In the Japanese sections, this zone is not sufficiently characterized by other fossil groups.

Notes. The index-species *Triassocampe deweveri* shows three pore sets in segments and no dorsal ridge. These characters suggest that this species could be a descendant of primitive *Triassocampe coronata* or *T. diordinia* (Bragin, 1999) that must be verified in the future.

There are minor differences between the late Anisian radiolarian assemblages of the Pacific and Tethyan regions. Abundant and diverse *Triassocampe* forms also appeared in the Tethys, but the genus *Yehararaia* is represented there by species other than those from the Pacific belt (Kozur and Mostler, 1994; Kozur *et al.*, 1996a).

Distribution: Sikhote-Alin', the Koryak Upland, Japan (the lower part of the analogous zone, according to Yao 1982; Matsuoka *et al.*, 1994), and Mediterranean region (Kozur and Mostler, 1994; Kozur *et al.*, 1996a).

UPPERMOST ANISIAN–LOWERMOST LADINIAN

Triassocampe scalaris Zone

Index-species: Triassocampe scalaris (Dumitrica, Kozur et Mostler, 1980, p. 26, Plate 9, Figs. 5, 6, 11).

Type area: Sikhote-Alin', the Dal'negorsk Settlement Bed 9, the lower part (Bragin, *et al.*, 1988; Bragin, 1991, with modifications).

Lithology: light gray to gray flaggy, locally massive glassy quartzite-like cherts. The zone spans the interval of 0–1.2 m from the base of the bed (Figs. 2, 3).

Nomenclature. The zone is newly established.

Stratigraphy and paleontological characteristics. The lower boundary is drawn at the first occurrence level of the index-species in association with *Spongostylkus tricostatus* Kozur et Mostler, *Spongolophophaena longa* Kozur et Mostler, *Welirella fragilis* Bragin, and *Yeharaia japonica* Nakaseko et Nishimura (Fig. 3). *Hozmadia rotunda* passes from the preceding zone. *Triassocampe coronata* disappears at the upper boundary of the zone, and *T. diordinis* passes into overlying deposits. The zone contains *Eptingium manfredi manfredi* (the species epibole is the lower Ladinian) and *Hindeosphaera spinulosa.*

Age substantiation. In the Dal'negorsk sections, this stratigraphic interval yielded early Ladinian conodonts *Neogondolella haslashensis* and *N. bakalovi* (Bragin, 1991). In the eastern shore of the Tonino-Aniva Peninsula (Sakhalin), the radiolarian assemblage of the zone is associated with the conodont species *Neogondolella transita* (the uppermost Anisian–lower Ladinian). Mediterranean analogues of the zone are characterized by ammonoids and conodonts characteristic of different levels of the upper Anisian and lower Ladinian (Dumitrica *et al.*, 1980; Kozur and Mostler, 1994; Kozur *et al.*, 1996a, 1996b). The lower boundary of the *T. scalaris* Zone is correlated with that of the *Tetraspinocyrtis laevis* Zone of the Mediterranean area (Kozur and Mostler, 1994).

Notes. The index-species *Triassocampe scalaris* has the well-sculptured test segments with distinct dorsal ridge thus continuing the phylogenetic line of the genus *Triassocampe* (Bragin, 1999).

Distribution: Sikhote-Alin' (the Dal'negorsk and Khabarovsk areas), Sakhalin (the Tonino-Aniva Peninsula), the Koryak Upland (the Ekonai Terrane and the Upper Khatyrka Melange) (Bragin, 1991), Japan (the Mino-Tamba and Chichibu terranes) (Yao, 1982, 1990; Matsuoka *et al.*, 1994), the Philippines (Yeh, 1990), Northern Italy (Dumitrica *et al.*, 1980; Kozur and Mostler, 1994), Southern Italy (De Wever *et al.*, 1990), and Hungary (Kozur and Mostler, 1994).

BASAL LOWER LADINIAN

Oertlispongus inaequispinosus Zone

Index-species: Oertlispongus inaequispinosus (Dumitrica, Kozur and Mostler, 1980, p. 5, Plate 10, Fig. 7).

Type area: Italy, the Passo della Gabiola section.

Nomenclature. The unit was first established as a middle subzone of the lower Ladinian (Kozur and Mostler, 1994). In the present article, it is ranked as a zone corresponding in scope to the Mediterranean *Oertlispongus primitivus* and *O. inaequispinous* subzones (Kozur and Mostler, 1994).

Stratigraphy and paleontological characteristics. The lower boundary is drawn at the first occurrence level of the index-species and associated *Pentaspongodiscus mesotriassicus* Dumitrica, Kozur et Mostler and *Spongostylus* sp. aff. *S. nakasekoi* Kozur et Mostler (Fig. 3). *Triassocampe coronata* and *T. deweveri* have their last occurrences at this boundary.

Age substantiation. In the type section, this stratigraphic interval corresponds to the upper part of the ammonite *Xenoprotrachyceras reitzi* Zone of the middle lower Ladinian (Kozur and Mostler, 1994). In the Dal'negorsk section, the unit is characterized by the early Ladinian conodonts (Bragin, 1991).

Notes. In addition to nassellarians, the family Oertlispongiidae appears to be of a great significance for the lower Ladinian stratigraphy. The family includes typical forms of Mesozoic spumellarians having spongy tests and complicated spines (genera *Oertlispongus* and *Falcispongus*, which substitute the primitive genus *Paroertispongus* and are succeeded by advanced *Spongoserrula* and *Pterospongus* genera). The stratigraphic range of Oertlispongiidae can serve as a base for more refined subdivision of the Ladinian Stage than it was done before (Bragin, 1999). However, such a progress is hampered by a relative scarcity of Oertlispongiidae in the Pacific region in comparison with the Mediterranean area (Bragin, 1994).

Distribution: northern Italy, Hungary (Kozur and Mostler, 1994), and Sikhote-Alin'.

TERMINAL LOWER LADINIAN AND BASAL UPPER LADINIAN

Falcispongus falciformis Zone

Index-species: Falcispongus falciformis (Dumitrica, 1982, p. 66, Plate 1, Fig. 5).

Type area: the Dal'negorsk section, Bed 9 (Bragin *et al.*, 1988), above the *Oertlispongus inaequispinosus* Zone.

Lithology: light gray to gray flaggy, locally massive glassy, quartzite-like cherts. The zone embraces the interval of 2.5–6 m above the bed base (Figs. 2, 3).

Nomenclature. The zone is newly established.

Stratigraphy and paleontological characteristics. The lower boundary of the zone is drawn at the first occurrence level of the index-species and associated *Pylostephanidium clavator* Dumitrica, *Yeharaia elegans* Nakaseko et Nishimura, and *Y. (?) annulata* Nakaseko et Nishimura (Fig. 3). Some species of the preceding zone, such as *Spongostylus tricostatus and Spongolophophaena longa*, disappear within the unit, and but others (*Eptingium manfredi, Pentactinocarpus awaensis* and *Pseudostylosphaera coccostyla*) become reduced in abundance.

Age substantiation. In the Dal'negorsk section, this interval contains the early Ladinian conodonts, being overlain by deposits attributed to the middle and upper parts of the upper Ladinian (Bragin, 1991). The lower boundary of the *Falcispongus falciformis* Zone corresponds to that of the *Ladinocampe multiperforata* Zone of the Mediterranean area (Kozur and Mostler, 1994). The mentioned zones are correlative with the upper part of the *Yeharaia elegans* Zone of Japan (Sugiyama, 1997).

Distribution: Sikhote-Alin', the Dal'negorsk area, and probably, the Mediterranean region.

MIDDLE AND UPPER PARTS OF THE UPPER LADINIAN SUBSTAGE

Muelleritortis cochleata Zone

Index-species: Muelleritortis cochleata (Nakaseko et Nishimura) (=*Emiluvia cochleata* Nakaseko et Nishimura, 1979, p. 70, Plate III, Figs. 2–4, 6).

Type area, Hungary, Balaton Highland, the Koveskal section (Kozur and Mostler, 1994).

Nomenclature. The zone was established by Kozur and Mostler (1994) in the scope of the middle–upper Langobardian. This interval was referred earlier to the *Sarla dispiralis* Zone of the upper Ladinian–Carnian (Bragin, 1991). The last taxon identified now as *Tritortis kretaensis dispiralis* (Bragin) is shown to be the most conservative representative of Muelleritortiidae. Therefore, it seems expedient to reject the *dispiralis* Zone in favor of more detailed subdivisions distinguished by Kozur and Mostler (1994).

Stratigraphy and paleontological characteristics. The lower boundary is drawn at the first occurrence level of the index-species and its subspecies (Kozur and Mostler, 1994) associated with *Hungarosaturnalis piletus* (Nakaseko et Nishimura), *Falcispongus curvispinosus* (Dumitrica), *Pseudostylosphaera longispinosa* Kozur et Mostler, *Silicarmiger latus* Kozur et Mostler, *Triassocampe* sp. ex gr. *T. sulovensis* Kozur, and *Tritortis kretaensis dispiralis* (Bragin). Simultaneously, this is the last occurrence level of *Falcispongus calcaneum, Oertlispongus inaequispinosus, Pseudostylosphaera japonica, P. coccostyla*, and *Hindeosphaera spinulosa* (Fig. 3). *Spongoserrula rarauana* Dumitrica appears inside the zone, whereas *Triassocampe scalaris, Silicarmiger costatus, Pylostephanidium clavator*, and *Pentactinocarpus fusiformis* still occur in this interval.

Age substantiation. In the Sikhote-Alin' section, this stratigraphic interval corresponds to the conodont *Sephardiella mungoensis* Zone (middle and upper parts of the upper Ladinian). Conodonts of the same age were found in the *Muelleritortis cochleata* Zone of Hungary (Kozur and Mostler, 1994). The same zone was also established in Japan (Sugiyama, 1997). The lower boundary of the *M. cochleata* Zone is one of the most reliable markers in the radiolarian stratigraphy.

Notes. The upper Ladinian is the first interval of the great Triassic extinction of radiolarians. Most of conservative taxa of Paleozoic and Early Triassic origin became extinct in this interval. *Triassocampe* forms of

reduced diversity become simple in morphology (*Triassocampe* sp. ex gr. *T. sulovensis*). Typical Mesozoic taxa, e.g., Muelleritortiidae, attain increasing importance. Family Oertlispongiidae gives rise to new genera *Spongoserrula*, and *Pterospongus* having main spines of complicated morphology. However, these taxa occur in abundance only in the Mediterranean region (Kozur and Mostler, 1996). In their vertical succession, radiolarians reveal distinct step-like patterns of appearance and extinction thus suggesting the short-term taxonomic renewals in composition of assemblages.

Distribution: Mediterranean region (Hungary) (Kozur, 1988a, 1988b; Kozur and Mostler, 1994), Sikhote-Alin, Sakhalin, the Koryak Upland (Bragin, 1991), Japan (Sugiyama, 1997), and Turkey (Bragin and Tekin, 1995).

LOWER CARNIAN

Tritortis kretaensis kretaensis Zone

Index-subspecies: Tritortis kretaensis kretaensis (Kozur)(=*Sarla kretaensis* Kozur, Krahl, 1984).

Type area: northern Hungary, the Dallapuzhta section (Kozur and Mostler, 1994).

Nomenclature. The zone was recognized by Kozur and Mostler (1994). In Primor'e, this interval was previously correlated with an upper part of the *Sarla dispiralis* Zone (Bragin, 1991).

Stratigraphy and paleontological characteristics. The lower boundary is drawn at the first occurrence level of the index-subspecies and associated *Pseudostylosphaera nazarovi* Kozur et Mostler, *P. inaequata* (Bragin), *P. goestlingensis* Kozur et Mostler, *Triassocampe sulovensis* Kozur, and *T. baldii* Kozur (Fig. 3). *Tritortis ariana* (Cordey), *Pseudostylosphaera imperspicus* (Bragin), *Xiphotheca (?) dimidiata* Bragin appear, and *Tritortis kretaensis dispiralis* continue to occur in the zone, whereas *Muelleritortis cochleata, Silicarmiger costatus S. latus*, and *Pylostephanidium clavator* become extinct.

Age substantiation. This stratigraphic interval yields the early Carnian conodonts *Neogondolella mostleri*, and *N. diebeli* in Hungary (Kozur and Mostler, 1994) and in Sikhote-Alin' sections.

Note. In Hungary, the lower boundary of the zone is placed at the level of sharp predominance of the index-subspecies over *Muelleritortis cochleata* (Kozur and Mostler, 1994). In the sections of Sikhote-Alin' and other eastern areas of Russia, the subspecies *Tritortis kretaensis kretaensis* has not been recorded below the lower Carnian.

Distribution: Hungary (Kozur and Mostler, 1994), Sikhote-Alin' (Dal'negorsk, Vysokogorsk), Sakhalin (Mt. Yunona) (Bragin, 1991), and Turkey (Bragin and Tekin, 1995).

UPPER CARNIAN

Capnuchosphaera theloides Zone

Index-species: Capnuchosphaera theloides De Wever (De Wever *et al.*, 1979, p. 84, Plate 4, Fig. 1).

Type area: Sikhote-Alin', the Dal'negorsk Settlement, Beds 10-12 (Bragin *et al.*, 1988; Bragin, 1991, with modifications).

Lithology: Bed 10 consists of gray–green massive diabases with inclusions of brownish green basalts and thin (0.5–1 m) interbeds and lenses of red occasionally bluish gray jaspers (7 m thick). Bed 11 of red, occasionally gray massive and indistinctly flaggy, frequently clayey jaspers is 2–4 m thick. Bed 12 is composed of dark greenish brown massive pillow-basalts with brownish to greenish red, frequently vesicular hyaloclastites between pillows, and with intercalations of bright red abiogenous jasperoids enclosing separate thin (0.5–2 m) interbeds and rare lenses of dark red indistinctly bedded jaspers. The bed is 15 m thick. (Figs. 2, 3).

Nomenclature. The zone is newly established. In scope, it corresponds to the previous *Capnuchosphaera lea* Subzone (Bragin, 1991). The easily identifiable species *Capnuchosphaera theloides* represents its index-species.

Stratigraphy and paleontological characteristics. The lower boundary of the zone is drawn at the first occurrence level of the index-species and other representatives of the genus *Capnuchosphaera*, such as *C. lea* De Wever, *C. tricornis* De Wever, *C. sagaris* Sugiyama, *C. triassica* De Wever, and others (Fig. 3). In addition, there appear many other taxa of the family Capnuchosphaeridae, e.g., *Icrioma tetrancistrum* de Wever and species of the genus *Sarla*. *Paleosaturnalis* shows its first occurrence in the zone, and *Japonocampe nova* (Yao), *Syringocapsa batodes* De Wever, and *Xiphotheca rugosa* Bragin appear among nassellarians. Last representatives of Oertlispongiidae and Muelleritortiidae completely disappear, but late species of the genus *Triassocampe* (*T. sulovensis*) continue to occur.

Note. The late Carnian radiolarian assemblage finally becomes dominated by taxa of the typical Mesozoic morphology. Radiolarians with skeletons of the initial Paleozoic type (families Pentactinocarpidae, Sepsagonidae, and others) are sporadic and not diverse.

At the base of the *Capnuchosphaera theloides* Zone in the Sikhote Alin', there may be a hiatus corresponding to the middle part of the Carnian Stage. In western Europe, this stratigraphic interval is referred to the *Tetraporobracchia haeckeli* Zone (Kozur and Mostler, 1994).

Age substantiation. In the Dal'negorsk section, the zone under consideration is characterized by late Carnian conodonts *Epigondolella abneptis* and *Neogondolella nodosa.*

Distribution: Sikhote-Alin', Sakhalin, the Koryak Upland (Bragin, 1991), Japan (Yao, 1982; Yoshida, 1986; Matsuoka *et al.*, 1994; Kamata, 1997; Sugiyama, 1997), the Philippines (Yeh, 1990). The Mediterranean *Nakasekoellus polita* Zone corresponds to the upper part of *C. theloides* Zone and contains similar radiolarian assemblage (Kozur and Mostler, 1994). In Japan, the given stratigraphic interval is subdivided to more details (Sugiyama, 1997; see the table).

LOWER–MIDDLE NORIAN

Capnodoce crystallina Zone

Index-species: Capnodoce crystallina Pessagno (Pessagno *et al.*, 1979, p. 176, Plate 1, Figs. 1–3).

Type area: Sikhote-Alin', the Dal'negorsk section, Bed 13 (Bragin *et al.*, 1988; Bragin, 1991 with supplements).

Nomenclature. The *Capnodoce antiqua* Subzone of the same scope was previously recognized in the type area (Bragin, 1991). The species *C. antiqua* is a junior synonym of *C. crystallina* (Sugiyama, 1997).

Lithology: gray, lilac–gray, less frequently light gray, thin- to medium-flaggy cherts with rare interbeds of yellowish gray siliceous mudstones. The zone embraces the interval of 0–6 m above the bed base (Figs. 2, 3).

Stratigraphy and paleontological characteristics. The lower boundary is drawn at the first occurrence level of the index-species and other representatives of the genus *Capnodoce*, e.g., of *C. crystallina* Pessagno, *C. ruesti* Kozur et Mostler, and *C. sarisa* De Wever (Fig. 3). In addition, *Nakasekoellus pessagnoi* (Nakaseko et Nishimura) and *Trialatus robustus* (Nakaseko et Nishimura) appear first at this boundary. Some characteristic species of the preceding zone (*Capnuchosphaera theloides* and *C. lea*) disappear. *Icrioma tetrancistrum, Kahlerosphaera norica*, and *Japonocampe nova* pass into the following zone.

Age substantiation. In the sections of Sikhote-Alin' (Dal'negorsk and Krasnaya Rechka) and Koryak Upland (Podgornaya), this stratigraphic interval contains conodonts of the *Epigondolella abneptis* and *E. postera* zones of the lower–middle Norian. The same interval in Japan has analogous paleontological characteristics (Yao, 1982). In the Mediterranean region and western North America, the proper radiolarian assemblages are associated with early–middle Norian ammonoids and bivalves (Kozur and Mostler, 1981, 1994; Blome, 1984; Yeh, 1989). The Mediterranean *Capnodoce ruesti* Zone corresponds to the lower Norian (Kozur and Mostler, 1994), and its lower boundary coincides with that of the *C. crystallina* Zone (table).

Distribution: Sikhote-Alin', southern Sakhalin, the Koryak Upland. The unit analogues are recognized in Mexico (Pessagno *et al.*, 1979), Oregon (Blome, 1983, 1984), Japan (Yao, 1982; Yoshida, 1986; Matsuoka *et al.*, 1994), and Mediterranean region (Kozur and Mostler, 1994).

BASAL UPPER NORIAN

Lysemelas olbia Zone

Index-species: Lysemelas olbia (Sugiyama, 1997, p. 160, Figs. 28-1, 43-6-10b).

Type area: Japan, the Inuyama section (Sugiyama, 1997).

Nomenclature. The zone was established in Japan in scope of the lower part of the upper Norian (Sugiyama, 1997).

Stratigraphy and paleontological characteristics. The lower boundary of the zone is drawn at the first occurrence level of the index-species and associated *Pseudoheliodiscus finchi* Pessagno, *Paleosaturnalis* sp. aff. *P. harrisonensis* (Blome), *Bipedis durus* Sugiyama, *Dumitricaella (?) parva* Sugiyama, and *Sarla hadrecaena* (De Wever) (Fig. 3). The genus *Capnodoce* disappears at this boundary. *Trialatus robustus* and *Kahlerosphaera norica* have their last occurrences in the zone. The transitory species *Pentactinocarpus sevaticus* is characteristic.

Age substantiation. The zone is characterized by the late Norian conodonts *Epigondolella bidentata.*

Distribution: Japan (Sugiyama, 1997), Sikhote-Alin'.

TERMINAL UPPER NORIAN

Betraccium deweveri Zone

Index-species: Betraccium deweveri Pessagno et Blome (1980, p. 235, Plate VI, Fig. 1).

Type area: eastern Oregon, the Supplee-Izee area, section of the Rail Cabin Formation.

Nomenclature. The zone was initially established in Oregon in a rank of a subzone (Blome, 1984) and recognized in the Far East of Russia as the lower subzone of the *Canoptum triassicum* Zone (Bragin, 1991). Its rank was elevated later to a zone (Carter, 1993).

Stratigraphy and paleontological characteristics. The lower boundary of the zone is drawn at the first occurrence level of the index-species. At the same level there appear abundant representatives of genera *Pantanellium* and *Praemesosaturnalis*, as well as species *Ayrtonius elisabethae* Sugiyama, *Bipedis acrostylus* Bragin, *Braginella rudis* (Bragin), *Canoptum rhaeticum* (Yao), *Livarella validus* Yoshida, and *L. densiporata* (Kozur et Mostler) (Fig. 3).

Age substantiation. In the Far East sections of Russia (Dal'negorsk, the Koryak Upland), this zone corresponds to the conodont *Epigondolella bidentata* Zone of the upper Norian. In North America, the zone is dated on the basis of concurrent late Norian conodonts (Carter, 1993).

Notes. At present, this zone is best traceable in radiolarian stratigraphy despite the slight variations in taxonomic composition of zonal assemblages. For instance, the assemblage with *Betraccium deweveri* from Cyprus is impoverished owing to its neritic origin (Bragin and Krylov, 1996).

The *B. deweveri* Zone was earlier considered as spanning the entire upper Norian. The latest investigations in Japanese sections (Sugiyama, 1997) demonstrated, however, that its index-species and typical accompanying taxa appear above the lower boundary of the upper Norian.

Distribution: Oregon, the Rail Cabin Formation (Blome, 1983, 1984; Yeh, 1989); Mexico, the California Peninsula (Pessagno *et al.*, 1979); British Columbia and Alaska (Carter, 1993); the Koryak Upland (Podgornaya), Sakhalin (Mt. Yunona), and Sikhote-Alin' (Dal'negorsk, Khabarovsk) (Bragin, 1991); Japan (Yao, 1982; Yoshida, 1986; Matsuoka *et al.*, 1994); the Philippines (Yeh, 1990, 1992); New Zealand and Cyprus (Bragin and Krylov, 1996); Turkey (Bragin and Tekin, 1996).

LOWER RHAETIAN

Livarella densiporata Zone

Index-species: Livarella densiporata (Kozur and Mostler, 1981, p. 115, Plate 9, Figs. 1a, 1b).

Type area: Austria, Zlambachgraben (Kozur and Mostler, 1994).

Nomenclature. Previously this stratigraphic interval represented an upper part of the *Canoptum triassicum* Zone (Yao, 1982; Bragin, 1991) and was termed the *Livarella gifuensis* Subzone (Bragin, 1991). The species *Livarella densiporata* has a wider geographic range and, therefore, is preferable as the index-species.

Stratigraphy and paleontological characteristics. The lower boundary of the zone is drawn at the first occurrence level of *Risella tledoensis* Carter, *Orbiculiforma* aff. *multibrachiata* Carter, *Proparvicingula* sp. cf. *P. moniliformis* Carter, and *Torosaturnalis orientalis* Bragin (Fig. 3). *Betraccium deweveri* disappears in this interval, but *Bipedis acrostylus, Canoptum rhaeticum, Deflandrecyrtium nobense, Braginella rudis,* and *Livarella validus* are still occurring here.

Age substantiation. In the type area, the zone is characterized by Rhaetian conodonts *Misikella posthernsteini* (Kozur and Mostler, 1994). This species was also found in the Sikhote-Alin' sections (Bragin, 1991).

Distribution: Mediterranean region (Kozur and Mostler, 1994) and eastern Russia (Bragin, 1991). Analogues of this zone are recognized in British Columbia (Carter, 1993), Japan (Yoshida, 1996; Sugiyama, 1997), and the Philippines (Yeh, 1992).

UPPER RHAETIAN

Globolaxtorum tozeri Zone

Index-species: Globolaxtorum tozeri (Carter, 1993, p. 111, Plate 19, Figs. 14–16).

Type area: Canada, British Columbia, the Queen Charlotte Islands, sections of the Kunga Island and Kennekott Point.

Nomenclature. The zone was established in British Columbia in the scope of the upper Rhaetian (Carter, 1993).

Stratigraphy and paleontological characteristics. The lower boundary is drawn at the first occurrence level of the index-species and associated *Bistarkum (?) cylindratum* Carter and *Pantanellium skidegatense* Pessagno et Blome (Fig. 3). Many species of two underlying zones, e.g., *Praemesosaturnalis hexaspinosus* (Bragin), *P. multidentatus* (Kozur et Mostler), *P. gracilis* (Kozur et Mostler), *P. heilongjangensis* Yang et Mizutani, *P. aciferus* (Bragin), and *P. diminutus* (Bragin), disappear at this level. *Bipedis acrostylus, Canoptum rhaeticum,* and *Dreyericyrtium ithacanthum* continue to occur in the zone. The most of Rhaetian radiolarian genera become extinct at the upper zonal boundary.

Age substantiation. In the stratotype, the zone is characterized by late Rhaetian ammonites *Choristoceras rhaeticum* Guembel and *C. nobile* Mojsisovics (Carter, 1993).

Notes. The Rhaetian radiolarian assemblages are peculiar in composition. Sharply dominant here are typical Mesozoic groups, families Pantanelliidae and Saturnalidae among them, which evolve further in the Jurassic. Many taxa, e.g., *Globolaxtorum tozeri* and *Pseudohagiastrum (?) tasuense* Carter, show morphological similarity with Jurassic radiolarians. Rare species with Paleozoic-type skeletons (*Braginella rudis* and *Pentactinocarpus magnus*) can be considered as relics.

At the same time, the further evolution of Late Triassic radiolarians was prevented by the great extinction event at the end of the Rhaetian. At this level, all species of the genus *Livarella* and all Ferresiidae taxa (genera *Ferresium* and *Risella*) disappear together with the latest forms of Pentactinocarpidae, Eptingiidae and Capnuchosphaeridae, whereas families Pantanelliidae and Saturnalidae show the completely renewed composition (Carter, 1993; Bragin, 1994). In contrast to the Rhaetian radiolarians, their earliest Jurassic assemblages have uniform and poor taxonomic composition (Carter, 1993; Carter *et al.*, 1998).

Distribution: British Columbia (Carter, 1993), Sikhote-Alin', and Japan (Sugiyama, 1997).

CONCLUSION

Potential of radiolarians for elaboration of the detailed Triassic biostratigraphy is not exhausted by the

described scheme. Further perspectives in this direction lie in a search of new radiolarian localities, especially in the sections of virtually unstudied Induan and lower Olenekian deposits. The Middle–Upper Triassic deposits can be successfully subdivided on the base of phylogenetic data on rapidly evolving groups, especially of the genus *Triassocampe* and families Oertlispongiidae and Saturnalidae. At present it is difficult to estimate the evolutionary rates of Triassic radiolarians because many of the taxa found has not been described yet. However, a great deal of short-living species suggests that the rates could be much higher in the Triassic time than in the Jurassic and Cretaceous periods. The detailed subdivision of the Middle–Upper Triassic on the basis of radiolarians and steadiness of radiolarian biostratigraphic units allow us to use this paleontological group not only for concrete stratigraphic purposes (e.g., for dating of siliceous deposits), but also in wider aspects of the global and regional Triassic stratigraphy.

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Dynamics of Diversity and Evolutionary Trends of Jurassic Foraminifers

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Abstract—Composition of the Jurassic (Hettangian–Tithonian) foraminiferal communities in three paleozoochores: Tethyan, Boreal-Atlantic, and Arctic, is considered. Foraminiferal taxons of all ranks (from species to orders) are analyzed with an emphasis given to the generic level, which objectively reflects the macroevolutionary process in the geological record. The Triassic–Jurassic boundary is marked by the mass extinction of foraminifers (about 55 genera) that was particularly pronounced among miliolids (22), involutinids (11), rotaliids (7), and lagenids (4).[1] Five major stages of foraminiferal evolution in the Jurassic are outlined: (I) Rhaetian–Lias (including the Hettangian *Planorbis* Zone) stage marked by the crisis in the foraminiferal development; (II) initial Early Jurassic stage characterized by the low diversity in the Hettangian–initial Sinemurian and by a sharp diversity increase at the end of the Sinemurian, in the Pliensbachian (particularly, in the late substage), and Toarcian; (III) Middle Jurassic (pre-Callovian) stage displaying the crisis trend at the Toarcian–Aalenian boundary and the taxonomic diversity increase in the Bajocian–Bathonian assemblages, particularly in the Tethys; (IV) Callovian–Oxfordian bloom universally accompanied by an intense radiation; and (V) Kimmeridgian–Tithonian stabilization period. The most important event in the evolution of Jurassic foraminifers was the appearance of first planktonic genera in the Bajocian. The foraminiferal diversity dynamics was fluctuating against the general background of about two-fold increase in the taxa abundance during the period from the Hettangian to Tithonian (60 and 127 new genera, respectively, versus 243 known genera of Jurassic foraminifers).

Key words: **foraminifers, diversity, Jurassic, biotic events, crises, extinction events, paleozoochores, evolution.**

The Jurassic, which, according to Hallam (1978), terminated the ancient stages of the Earth evolution, was a period, when biota retaining the Early Mesozoic and, partly, Late Paleozoic peculiarities acquired simultaneously new features, which became preponderant in the Cretaceous and later epochs.

The development of marine biota in the Jurassic occurred against the background of significant events in the Earth history: termination of the early Cimmerian orogeny; opening of the Atlantic; destruction of the Tethyan carbonate platform; pre-Cretaceous late Cimmerian uplifts; major transgressions and regressions that changed dimensions and configuration of sea basins in the northern hemisphere; and climatic changes that resulted in increased differences between zoochores and, respectively, in elevation of their ranks and appearance of the Arctic paleozoogeographic province. Nonetheless, no mass extinctions or catastrophic events, which would eliminate the substantial part of biota, have occurred during the Jurassic period 70 m.y. long. To the contrary, as is evident from the foraminiferal evolution, many groups experienced the intense species radiation, which resulted in the appearance of new larger taxons (genera, families, orders), first planktonic forms included. This important stage in the foraminiferal evolution was to a large extent responsible for their later development. In this work, we consider the composition of foraminiferal assemblages and their changes during the Jurassic, as well as probable causes responsible for these transformations.

The analyzed material includes own samples collected from many Jurassic sections in different climatic zones during several decades and abundant published data.

As far back as in the 1970s, we defined, based on peculiarities in differentiation of the Jurassic foraminifers, three types of faunas or faunal groupings, which cannot be further enlarged and differ in composition, origination area, and formation history (Basov, 1974; Kuznetsova, 1979) (Fig. 1).

(1) The tropical, cyclamminid–pfenderinid fauna. This type includes abundant highly specialized endemic taxons (with an intricate cellular or alveolar inner structure), the so-called "larger" agglutinated foraminifers belonging to the Lituolida, Ataxophragmiida, and Orbitolinida orders and calcareous porcelain forms of the Miliolida order.

(2) The subtropical nodosariid–epistominid fauna. It is characterized by the absence of all the above-men-

[1] Shown in brackets are numbers of extinct foraminiferal genera.

Fig. 1. Geographic distribution of main foraminiferal fauna types in the Jurassic: (1) nodosariid–ammodiscid (moderately cold-resistant); (2) nodosariid–epistominid (subtropical); (3) cyclamminid–pfenderinid (tropical); (4) boundaries of paleozoochores; (5) contours of continents.

tioned highly specialized forms and shows preponderance of calcareous foraminiferal species having the porcelain and divergent structure of tests and displaying simple inner structures of chambers. These are mainly representatives of the cosmopolitan order Lagenida and rotaliid families Ceratobuliminidae and Epistominidae. The important role in the fauna of this type belongs to miliolids, spiriolinids, and involutinids widespread in the Tethyan province and missing, similarly to epistominids, from the Arctic region. Endemic taxons are represented at the generic level only.

(3) The temperate cold-resistant nodosariid–ammodiscid fauna. This group is dominated, along with calcareous cosmopolitan lagenids, by smaller agglutinated foraminifers (usually with the siliceous or ferrous cement). Preponderant are genera of orders Ammodiscida, and Nodosariida, as well as those belonging to the families Trochamminidae, Ataxophragmiidae, and Textulariidae. Endemism is high only at the species level.

The Triassic–Jurassic boundary was marked by the mass extinction of marine organisms, and this event was one of most significant in the Earth history (Hallam, 1995; Alekseev, 1998). The event influenced foraminifers as well: about 55 genera (35% of total 160 genera known from the Triassic) and also some families and subfamilies became extinct. The largest number of genera disappeared amid miliolids (22), involutinids (11), and rotaliids (7). The mass extinction of Triassic taxons made vacant previously occupied niches and favored the rapid evolution and distribution of newly formed and survived genera and species in the transgressive sea basins of the Jurassic.

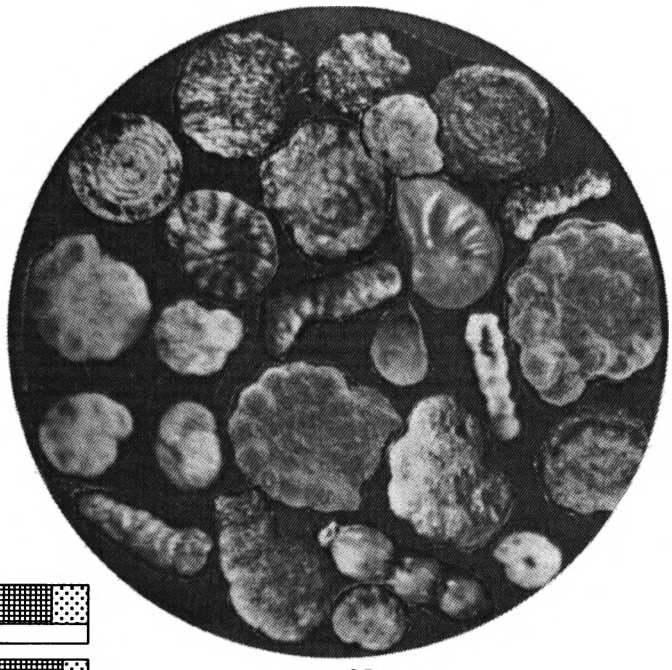
The Jurassic period was a time of active radiation, which resulted in appearance of new species (to 50 species of the single dominant genus *Lenticulina* per age) and genera (for instance, 24 new genera appeared equally during the Bajocian and Oxfordian). New taxons of a higher rank also appeared, namely, 33 families and 8 orders and suborders. The appearance of "larger" agglutinated lituoilids and orbitolinids together with origin of secretory calcareous ceratobuliminids, epistominids, buliminids, and, particularly, of planktonic globigerinids represent the biotic events of a high evolutionary importance, which determined the later development of foraminifers in the Cretaceous.

To gain an idea of the diversity of the Jurassic foraminiferal fauna, we compiled lists of species for every stage and every of three main paleozoochores (Arctic, Boreal-Atlantic, and Tethyan) using our own and many published taxonomic and stratigraphic data, first of all compendiums and paleontological atlases (Hottinger, 1967; Grigelis, 1985; Kuznetsova *et al.*, 1996). We revised as well old publications (e.g., those of Seibold and Seibold, 1956; Bach *et al.*, 1959; Ruget, 1976; and others).

The different state of knowledge and dissimilar quality of published data on foraminifers complicated this task. The most detailed studies are those on foraminifers from northern hemisphere, particularly, on their boreal communities. Tethyan foraminifers are less studied, whereas only fragmentary data are available for the Notal realm. The analysis of the global composition and distribution of Jurassic foraminifers made it necessary to use all published data, those concerning oceanic sediments recovered by deep-sea drilling inclusive (Luterbacher, 1972; Bartenstein, 1974; Kuznetsova, 1974; Kuznetsova and Seibold, 1977; Gradstein, 1983; Gradstein and Sheridan, 1983; Riegraf *et al.*, 1984).

It should be mentioned that the compilation of complete species lists for stages (the total list of Jurassic foraminifers is estimated to include 1000 species and more) is an extremely complicated task, which cannot satisfactorily be solved without the ultimate complete analysis of publications (in case of foraminifers, they are extremely abundant and often published in hardly accessible editions) and their thorough revision to avoid synonyms. From this standpoint, we are only halfway, and, consequently, the presented results should be considered as a first attempt, which provides not ultimate, but, in terms of statistics, the rather trustworthy picture of changes in the taxonomic diversity, at least at the generic level, if one takes into consideration an abundance of the source materials.

Keeping this in mind, we determined the abundance and generic composition of all 17 known foraminiferal



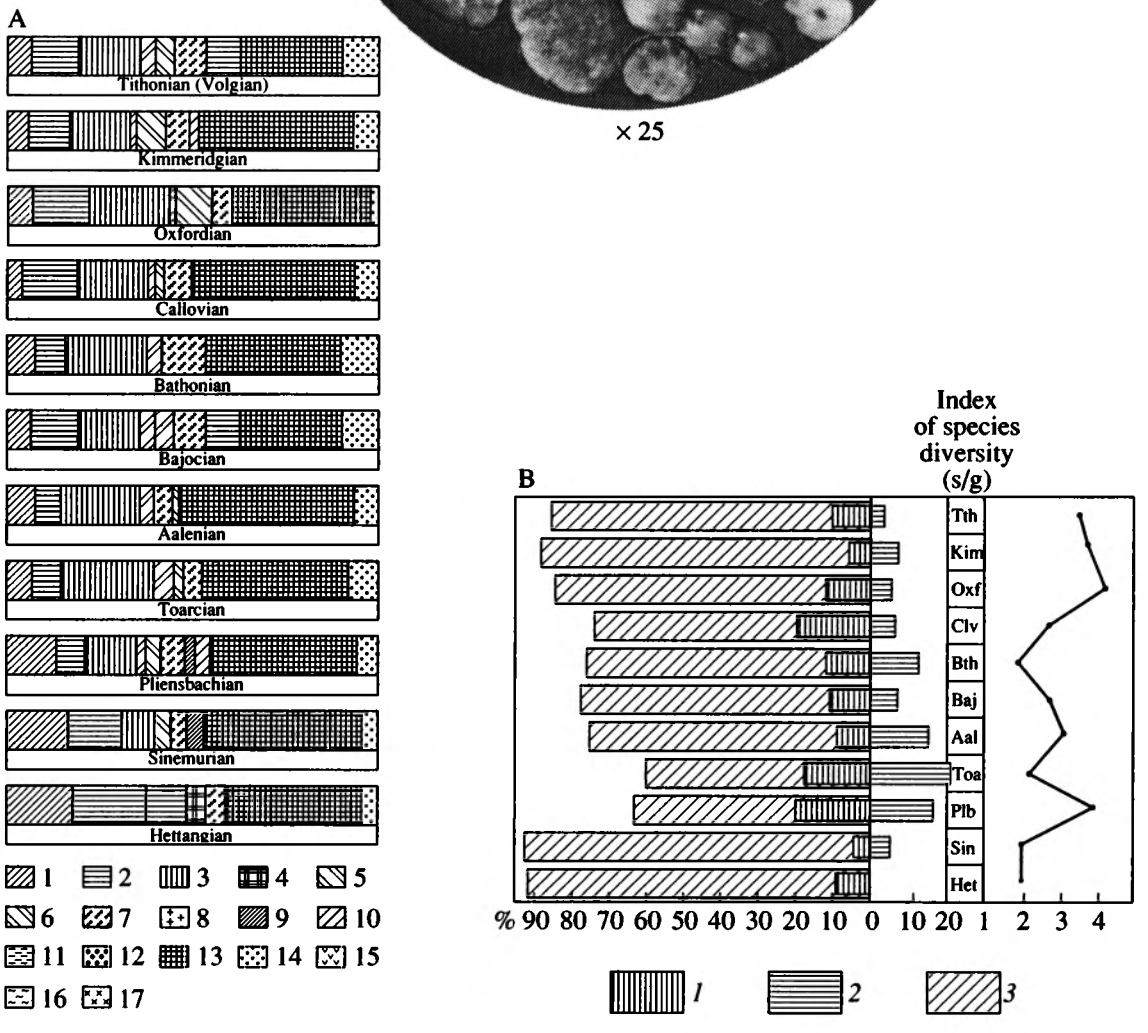


Fig. 2. (A) Taxonomic composition of Jurassic foraminiferal communities (orders, in %) in the Arctic zoochore. Foraminiferal orders: (1) Astrorhizida, (2) Ammodiscida, (3) Lituolida, (4) Rzhehakinida, (5) Trochamminida, (6) Textulariida, (7) Ataxophragmiida, (8) Orbitolinida, (9) Involutinida, (10) Cornuspirida, (11) Spirillinida, (12) Miliolida, (13) Lagenida, (14) Polymorphinida, (15) Rotaliida, (16) Globigerinida, (17) Buliminida, and (B) variations in abundance of appeared, extinct, and survived genera of Jurassic foraminifers (in %). (*1*) Appeared; (*2*) extinct; (*3*) survived genera.

orders for every Jurassic age using classification by Maslakova and Gorbachik (1995) with due account for classification by Loeblich and Tappan (1988).

The total taxonomic composition and structure of foraminiferal communities for main zoochores are shown in Figures 2A, 3A, and 4A. They demonstrate proportions (%) of representatives of all 17 orders calculated for every stage.

Considering the generic composition of successive foraminiferal communities in every geological period, we can subdivide them according to the appearance, disappearance, and survival. Presenting these parameters as percentage of the total number of genera recorded in the particular stage, we can obtain curves characterizing dynamics of the taxonomic diversity and evolutionary processes during the Jurassic (Figs. 2B, 3B, and 4B).

In this work, we used the following parameters.

(1) Index of the taxonomic diversity corresponding to the ratio between a number of low-order taxons and those of a higher order (species and genera, genera and families or orders, etc.) calculated for the particular period of geological time (age).

(2) Index of extinction characterizing the number of extinct genera expressed in percents relative the total generic diversity during the considered age.

(3) Rate of extinction or the ratio between the coefficient of extinction at the critical boundary and the background extinction level (during previous "quiet" ages).

(4) Index of disappearance rate characterizing in the above manner both the extinct genera and those disappeared for a certain period (Lazarus-taxons, see below).

(5) Index of appearance rate or the number of first appeared species expressed in percents of the total number of genera existing during the considered age.

(6) Index of renewal rate, which is applied as previous one to the first appeared taxons and those reappeared (Lazarus-taxons).

(7) The quotient of changes at the stratigraphic stage boundaries, which represents ratio (%) of the total number of genera appeared above and disappeared below the particular boundary to the total generic diversity characteristic of both stages.

The last parameter corresponds to the coefficient of changes in the taxonomic composition at stratigraphic boundaries proposed by Yu.E. Tkachev and used by Voronov (1994) for ammonoids:

$$R = \frac{N(H) + N(B)}{N(H) + N(B) + N(T)},$$

where N(A) is a number of taxons appeared at the considered stratigraphic boundary, N(E) is the same for extinct taxons, and N(T) is a number of transient taxons.

It is easy to note that in our case, N(A)+N(E)+N(T) is a sum of all genera registered in both stages, whereas the quotient of changes corresponds to the Tkachev's *coefficient* of changes expressed in percents. The term *quotient* is more suitable for our purpose than the term *coefficient*, because it allows understanding of a certain process development.

Discrimination of notions extinction and disappearance or appearance and renewal of taxons is necessary, as there are two groups amid "appearing" and "disappearing" taxons (in the considered case, genera), both equally important for understanding of evolutionary processes, namely, genera that first appeared at the considered level of the Earth history (for instance, *Tristix* in Pliensbachian, *Epistominoides* in Bajocian) and genera that existed in earlier epochs, were missed from several ages, and reappear in younger deposits (the so-called Lazarus-taxons). The above-mentioned *Tristix*, which was never registered in the Toarcian and Aalenian, appeared again in the Bajocian and survived through the remaining part of the Jurassic, Cretaceous, and Paleogene (probably being absent at some levels). *Epistominoides* is known to exist in the Bajocian to Kimmeridgian and, afterward, only in the Barremian and Paleogene (Grigelis, 1985). *Meandrospira*, the well-known genus from the Triassic deposits, is not recorded in the entire Jurassic succession, though it was sufficiently widespread in the Early Cretaceous beginning from the Valanginian. Such patterns of the genus distribution can be related to the incomplete paleontological record either due to the significant progressive reduction of the abundance and distribution area of the genus, or owing to the gap in the phylogenetic succession of forms referred to the considered genus. The latter case implies that either our knowledge about the taxon is inadequate and the younger forms bearing features of this genus should be referred to the new species, or genetic mechanisms that would allow temporal loss and subsequent restoration of features of a certain genus should be kept in mind.

In this connection, it is possible to calculate the number of appeared and extinct genera and, consequently, the index of the community renewal for every age, with or without account for the Lazarus-taxons, and the two results obtained will be different (Fig. 5).

In total, 243 foraminiferal genera (239 benthic and 4 planktonic taxons) were registered in the Jurassic deposits (Table 1). Of these taxons, 174 genera (or 72%) first appeared in the Jurassic. The average origination rate is 8–12 genera per age, but it could be twice greater during the period of intense radiation.

The initial Jurassic was characterized by the low taxonomic diversity, because foraminiferal communities were recovering after the mass extinction at the end of the Triassic (Table 2). Of an extremely low diversity is the foraminiferal assemblage of the *Planorbis* Beds, but the number of genera grew to 60 already at the end of the Pliensbachian, when 21 new Jurassic genera already appeared (36% of the total genera number). The subsequent gradual diversity increase reached the

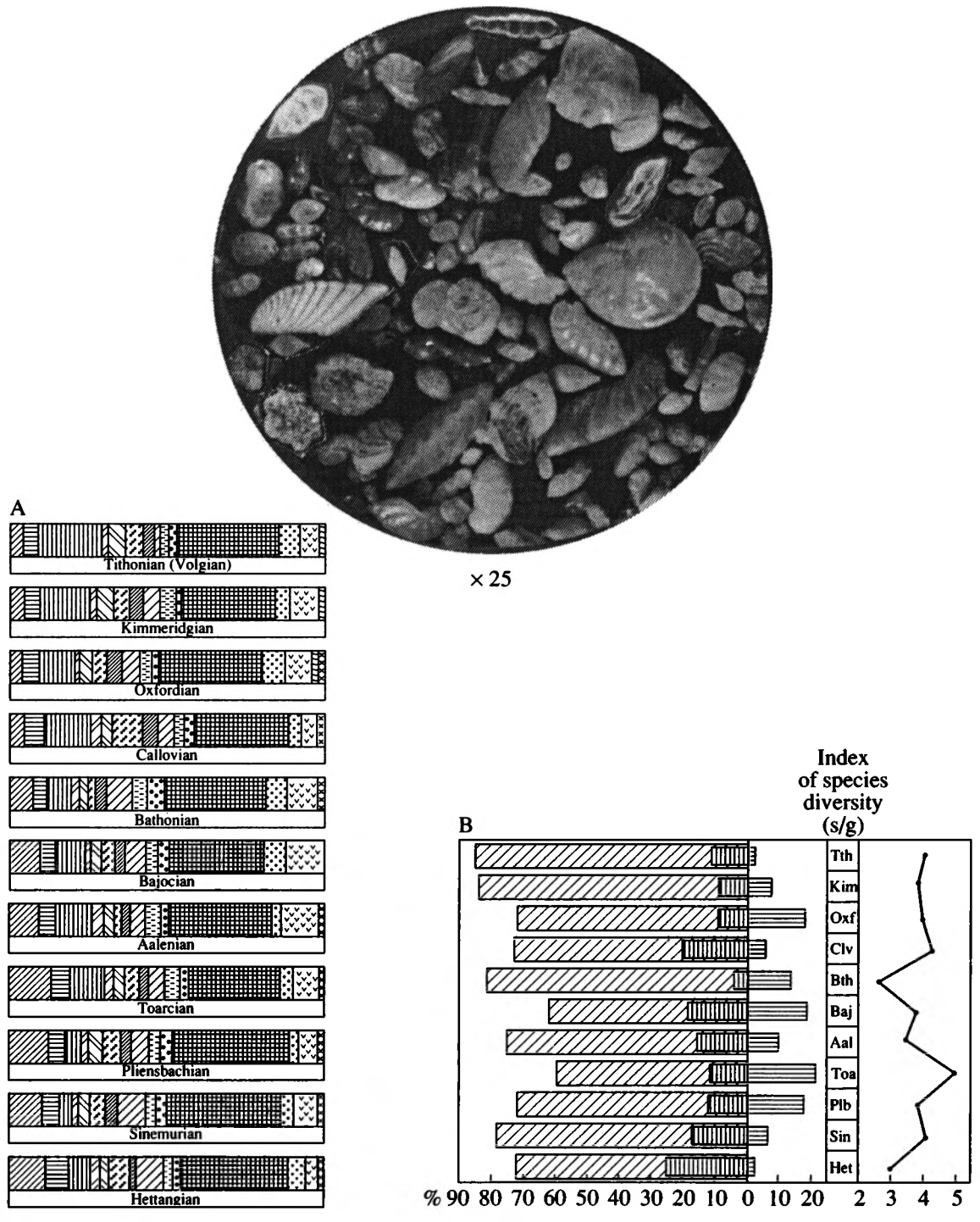


Fig. 3. (A) Taxonomic composition of the Jurassic foraminiferal communities (orders, in %) in the Boreal-Atlantic zoochore, and (B) variations in abundance of appeared, extinct, and survived genera of Jurassic foraminifers (symbols as in Fig. 2).

Early Jurassic maximum (95 genera) in the Pliensbachian. The Early Jurassic was the appearance time of 16 new families (48% of new Jurassic families) and 3 new orders and suborders. In the post-Pliensbachian time (Toarcian and Aalenian), there was a certain reduction of species and generic diversity (to 73 genera) followed by the new rapid increase to 117 genera in the Bajocian, and this trend continued in the Bathonian and Callovian and reached the maximum of 144 genera in the Oxfordian. The terminal Jurassic was marked by stabilization of the genera diversity at the level of 135 genera and by slight reduction of species diversity. Noteworthy is also

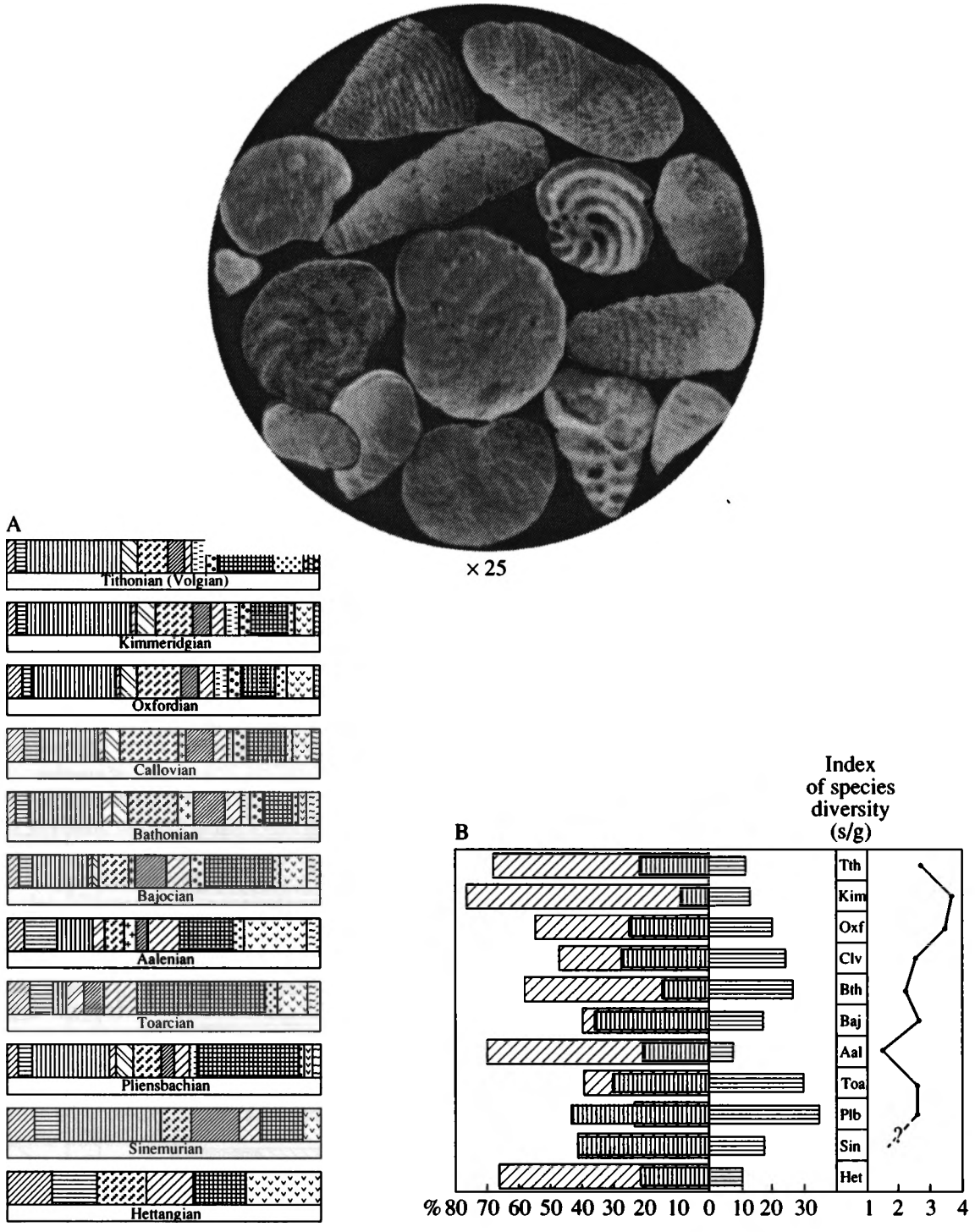


Fig. 4. (A) Taxonomic composition of the Jurassic foraminiferal communities (orders, in %) in the Tethyan zoochore, and (B) variations in abundance of appeared, extinct, and survived genera of Jurassic foraminifers (symbols as in Fig. 2).

origination dynamics of new higher-rank taxons: 12 new families and 4 orders and suborders originated in the Middle Jurassic and only 5 families (3 of them in the Oxfordian) and 1 suborder appeared in the Late Jurassic.

It should be noted that the Toarcian–Aalenian crisis particularly distinct in the reduced taxonomic diversity of Tethyan and Arctic foraminiferal faunas began with the temporal extinction of many Jurassic genera probably due to reduction of distribution areas and retreat of

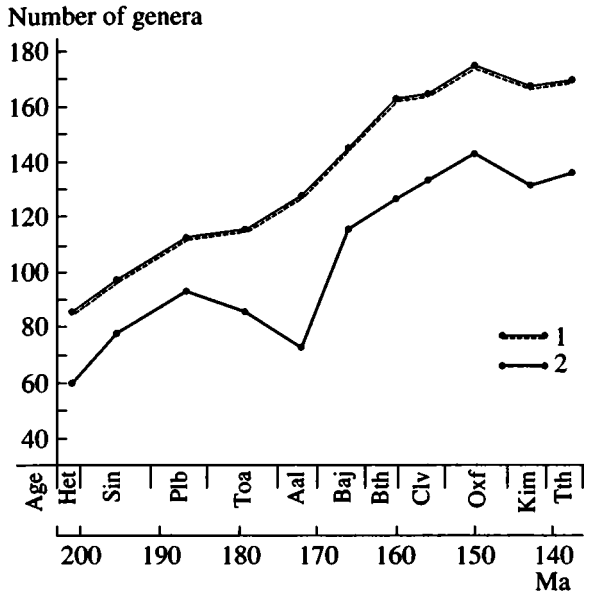


Fig. 5. Generic diversity of Jurassic foraminifers: (1) number of genera, Lazarus-taxons included; (2) number of genera, Lazarus-taxons excluded.

respective taxons into some isolated refugia to survive unfavorable conditions. This behavior resulted in their absence in sections (an apparent extinction).

Of interest are the data on the species diversity obtained for the above time interval (Fig. 6). The indices of species diversity calculated for zoochores turned out to be in antiphase: the Toarcian maximum of species diversity in the Boreal Atlantic zoochore corresponds to its reduction in the Arctic regions, and the Aalenian minimum in the Tethyan zoochore was concurrent to reduction in the Boreal Atlantic and to simultaneous increase in the Arctic zoochore. All other ages of the Jurassic are characterized by the more or less consistent changes in the indices of species diversity in different zoochores (Table 3).

All this suggests most likely an intense migration of foraminifers during the Toarcian–Aalenian, their escape from unfavorable conditions that existed in some zoochores and were related, for instance, to the well-known early Toarcian anoxic event.

Therefore, it is not reasonable to consider the Toarcian–Aalenian as a period of remarkable mass extinction, although some researchers deem extinction of ammonites at the Pliensbachian–Toarcian boundary as representing such a crisis.

The predictive post-crisis growth of the taxonomic diversity occurred in the Bajocian, but it is notable only in the Tethyan and Boreal Atlantic zoochores (Table 2). In the Arctic zoochore, there was some reduction of foraminiferal diversity as compared with that in the Aalenian that was related to isolation of Arctic seas. The last

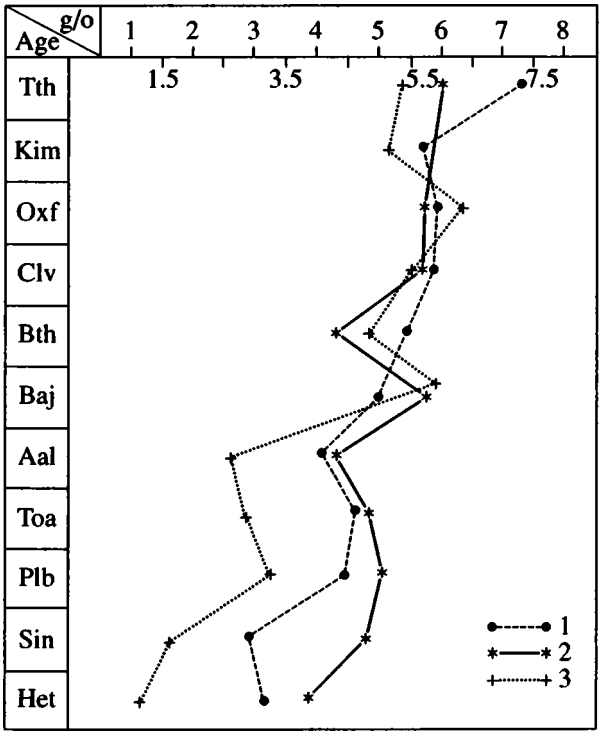


Fig. 6. Variations of the generic diversity index (g/o) for Jurassic foraminifers. Zoochores: (1) Arctic, (2) Boreal-Atlantic; (3) Tethyan.

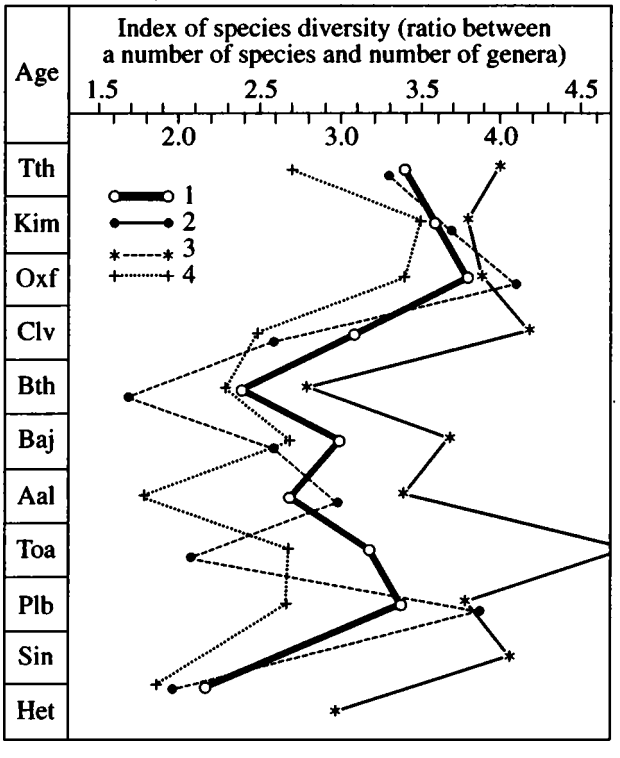
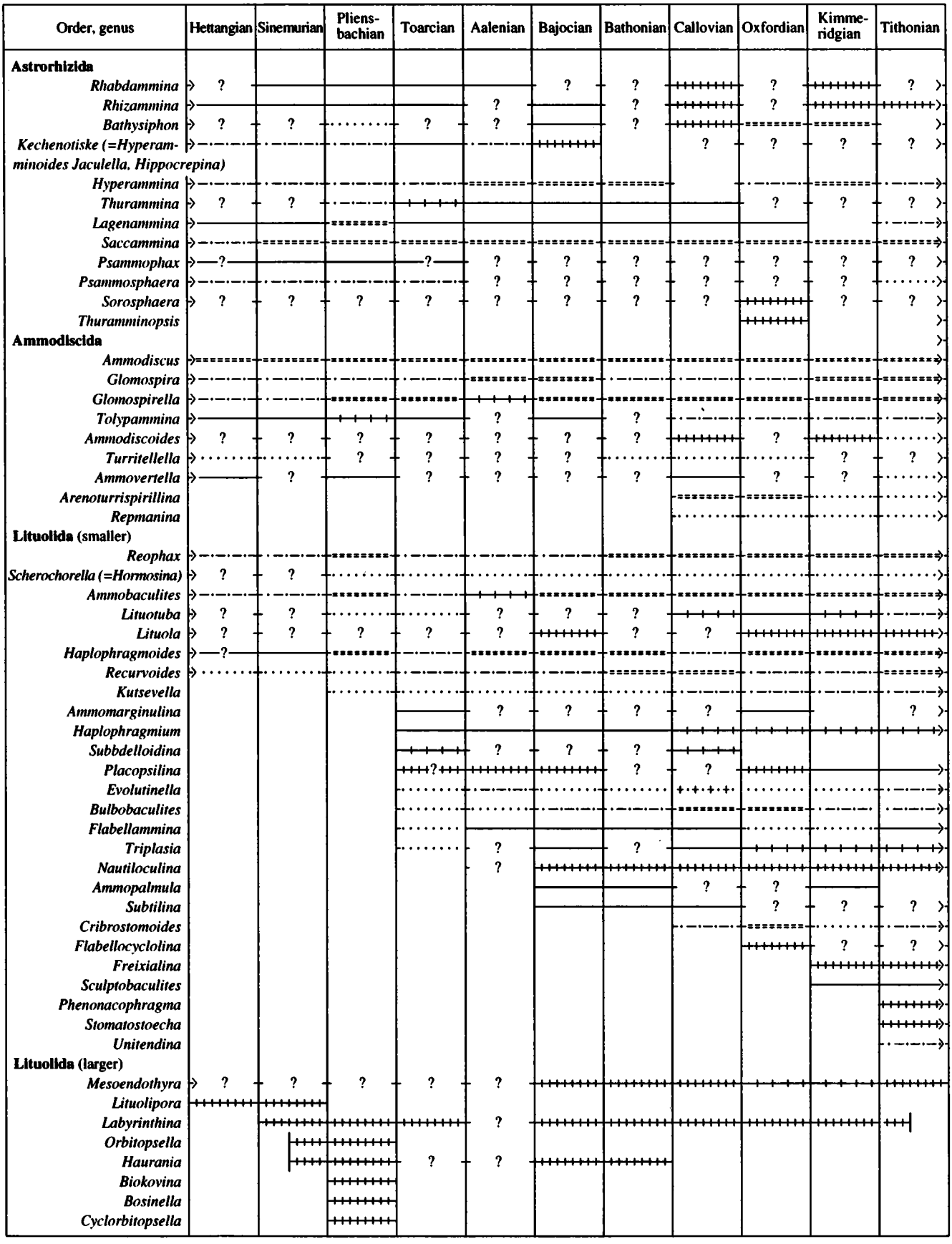


Fig. 7. Variations of the species diversity index (s/g) for Jurassic foraminifers. (1) Average variations of the species diversity index (s/g) for Jurassic foraminifers Zoochores, (2) Arctic, (3) Boreal-Atlantic, (4) Tethyan.

Table 1. Stratigraphic and paleobiogeographic distribution of Jurassic foraminifers



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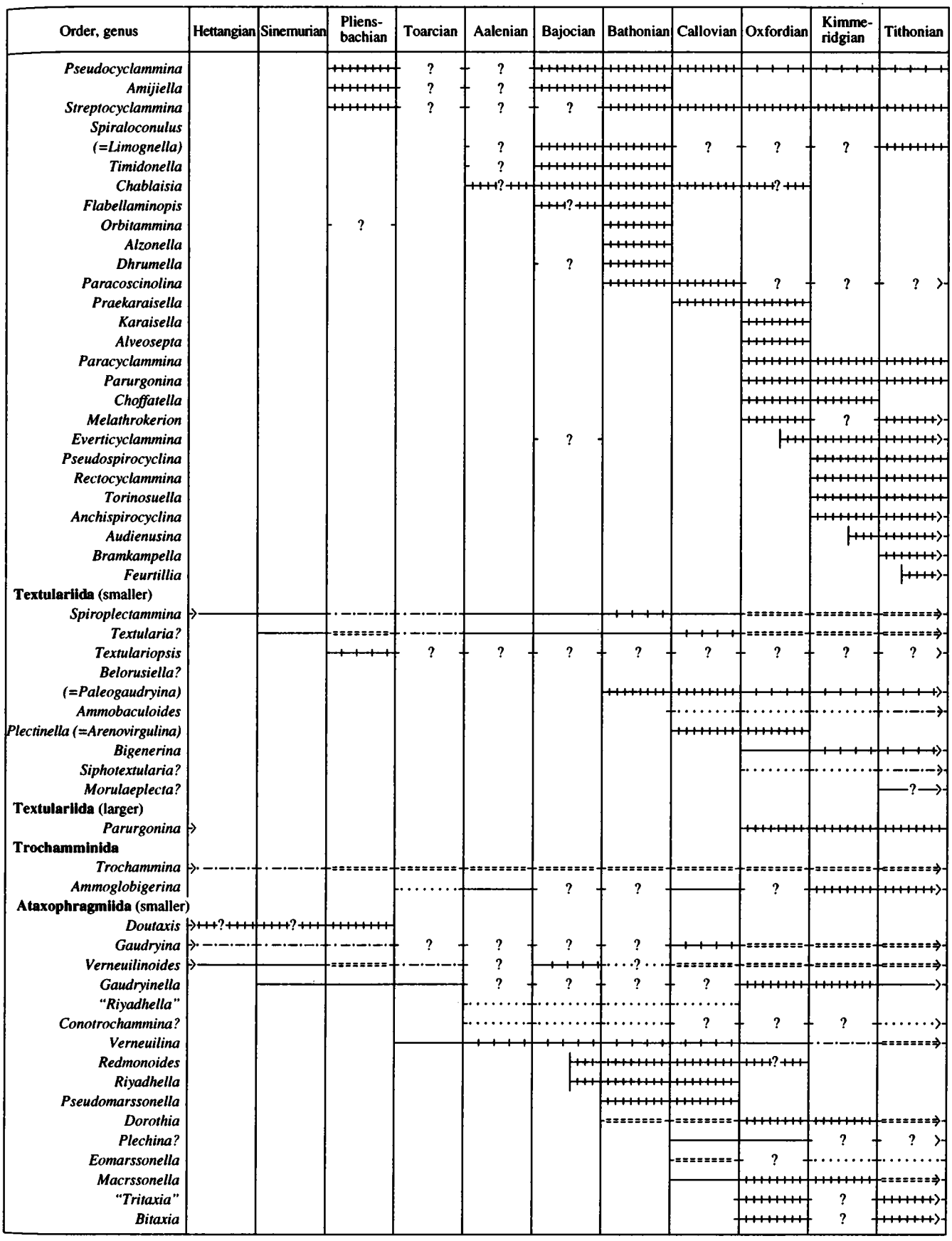

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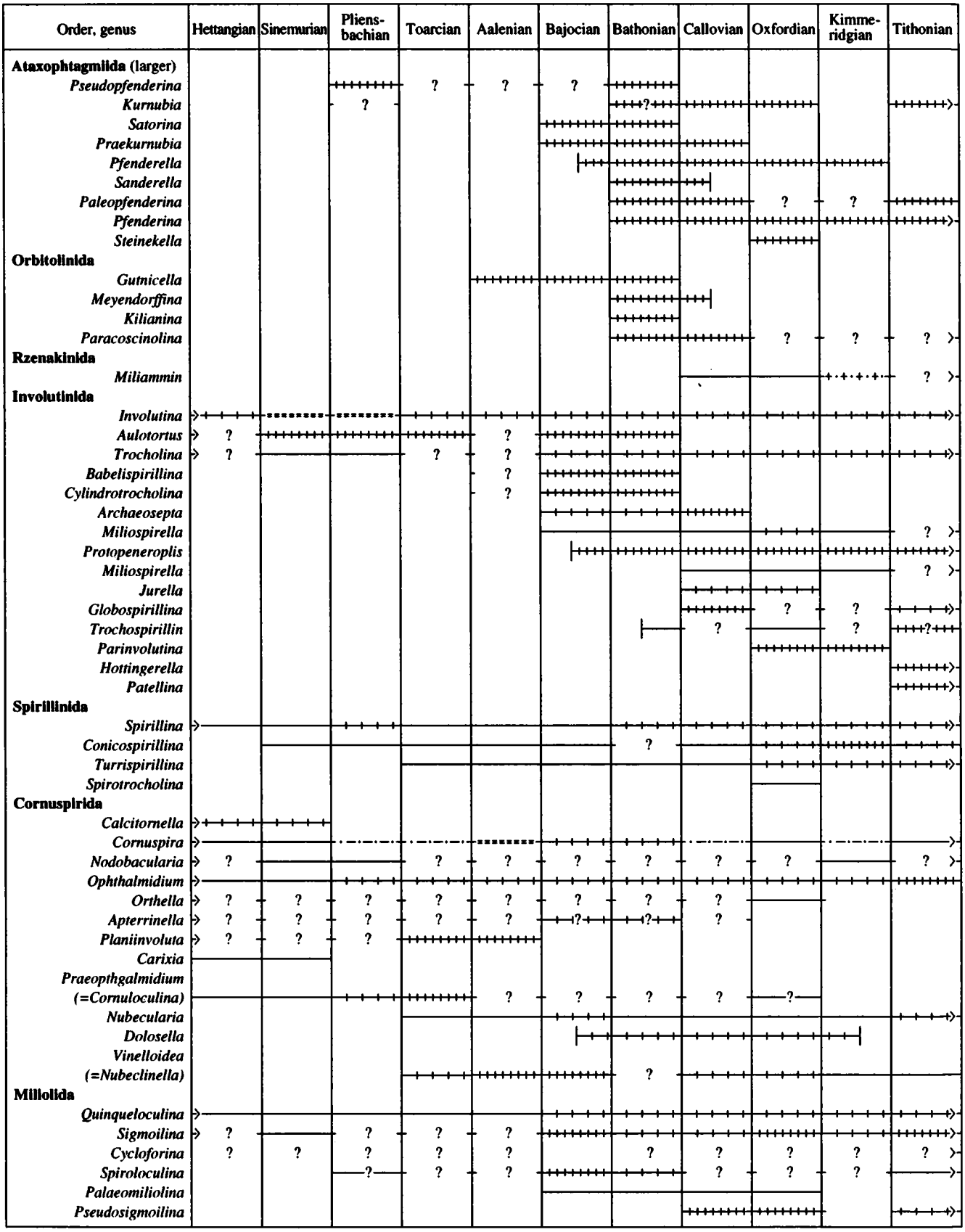

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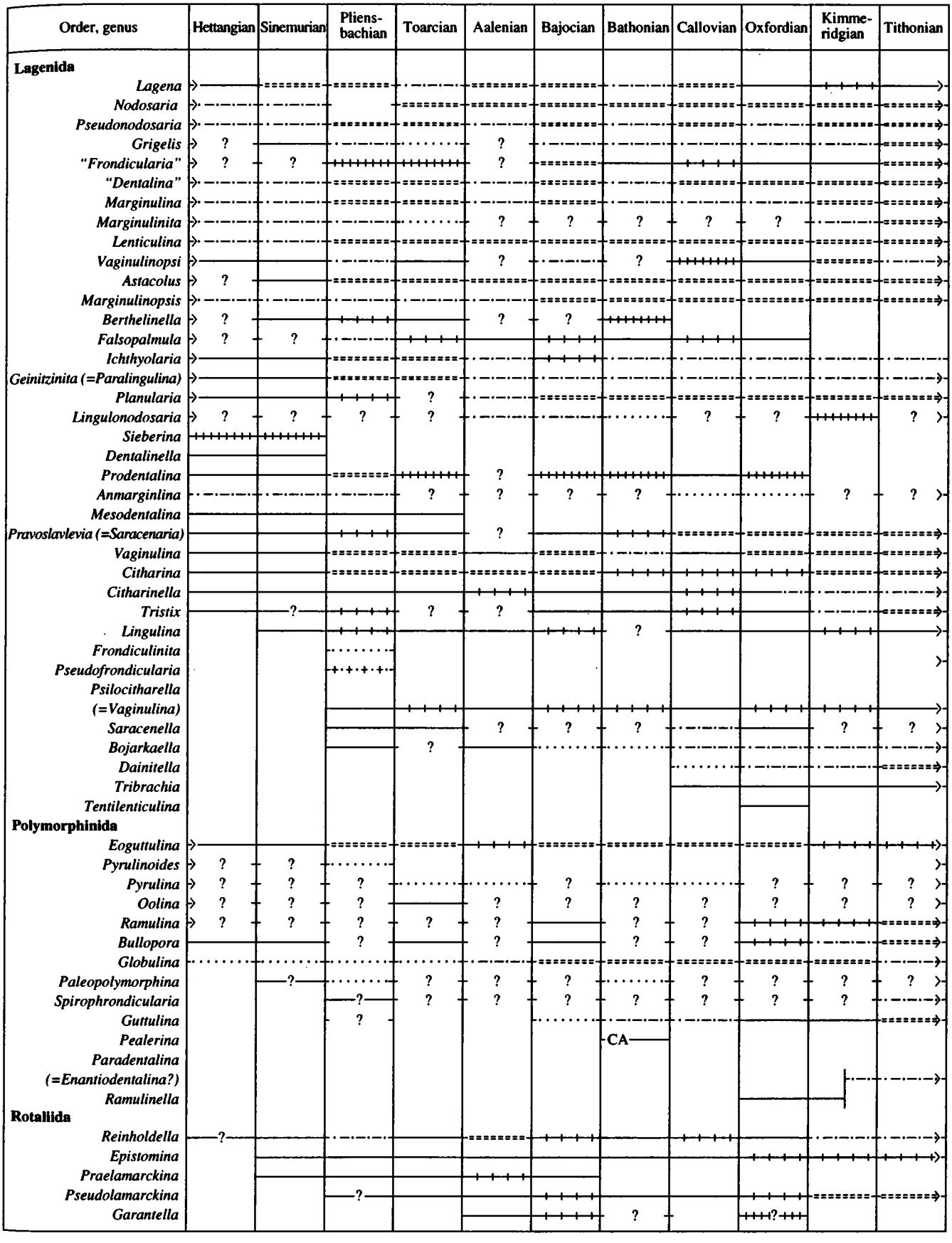

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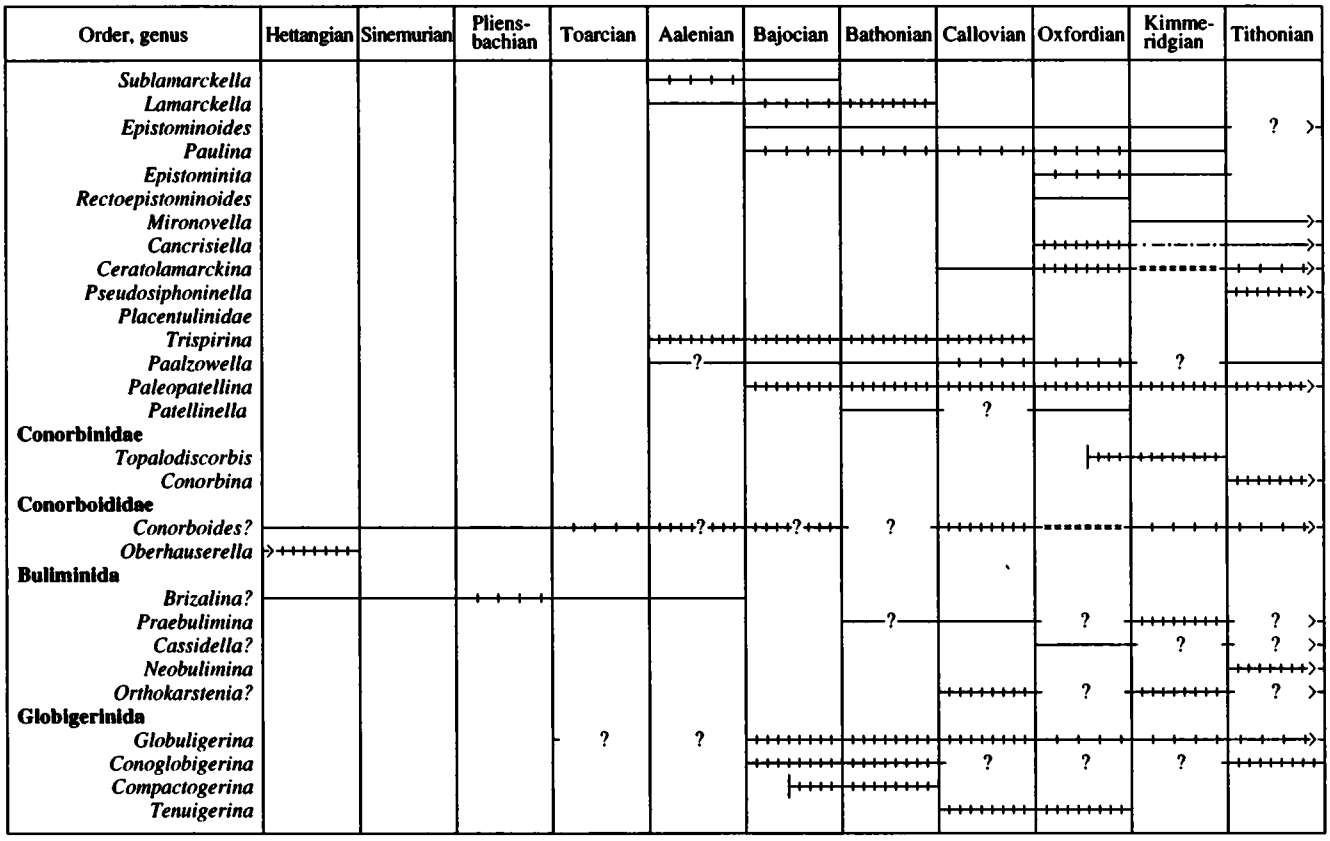

Total: 243 genera, 174 of which appeared in the Jurassic (69 transient genera)

Zoochores:

+++++ - Tethyan, ——— - Boreal-Atlantic, ····· - Arctic,

+-+-+ - genera distributed in two contiguous zoochores (semicosmopolitan),

▪▪▪▪▪▪ - cosmopolitan genera, (NA) in North America,

⊢› ›⊣ - genera originated in the Late Triassic and continuing to exist in the Cretaceous.

Table 2. Number of foraminiferal genera in the Jurassic zoochores

Epoch	Age	Zoochore			Total	Including		AED*
		1	2	3		new	in zoochores	
							1 2 3	
J_3	Tithonian (Volgian)	80	92	64	137	11	9–2–0	1
	Kimmeridgian	67	60	52	133	9	7–2–0	1
	Oxfordian	70	77	44	144	[24]	15–5–1	1
J_2	Callovian	73	68	42	135	[19]	9–7–3	1, 2
	Bathonian	73	28	36	126	[18]	13–4–1	1
	Bajocian	59	58	30	117	[25]	19–5–1	1
	Aalenian	18	50	28	73	9	4–3–2	1, 2
J_1	Toarcian	26	67	36	86	13	3–5–5	2, 3
	Pliensbachian	36	59	42	94	[17]	9–5–3	1, 2
	Sinemurian	12	51	23	77	11	3–8–0	2
	Hettangian	7	39	17	60	[17]	2–14–1	2
					Total:	173	96–60–17	

(AED*) Areas with elevated diversity (numbers in rectangles denote maximums).
Zoochore: (1) Tethyan, (2) Boreal-Atlantic, (3) Arctic.

Table 3. Species diversity of Jurassic foraminifers

Age	Zoochore*			Average
	1	2	3	
Tth	2.67	4.02	3.38	3.35
Kim	3.50	3.80	3.69	3.66
Oxf	3.38	3.85	4.14	3.79
Clv	2.54	4.21	2.61	3.12
Bth	2.40	2.81	1.75	2.32
Baj	2.60	3.73	2.60	2.97
Aal	1.55	3.38	3.03	2.65
Toa	2.61	4.89	2.15	3.21
Plb	2.60	3.82	3.88	3.43
Sin	1.35	4.16	1.90	2.27
Het		3.00		

* Zoochore: (1) Tethyan, (2) Boreal-Atlantic, (3) Arctic.

event lowered the water temperature and increased originality in other faunal groups, thus allowing experts on the molluscan fauna to consider the Arctic zoochore as a higher-rank biogeographic unit (province) of the Bajocian and subsequent periods (Saks *et al.*, 1971).

The Bathonian age marks a new reduction of species diversity (Fig. 6), while the number of foraminiferal genera remained approximately equal to that in the Bajocian. Moreover, the Tethyan seas were characterized by the noticeable diversity growth mainly owing to endemics, such as "larger" agglutinated foraminifers that occupied carbonate platforms and reefs (Table 2, Fig. 7).

The peak of the generic and species diversity coincides with the Oxfordian (Figs. 4 and 7). The post-Oxfordian period was marked by stabilization of taxonomic diversity at the lower level (about 130 genera) because of the elevated extinction rate at the end of the Oxfordian. In addition, new high-rank taxons of this period are extremely rare.

Table 4 demonstrates all data on the community renewal rate, evolutionary appearance of new genera, and their extinction. Against the background of the general growth of the taxonomic diversity, there were periods of the sharp increase in the extinction (disappearance) rate like the terminal Pliensbachian and Bathonian (to a lesser extent, Oxfordian) followed by the subsequent enhanced renewal rate (elevated quotient of changes).

Three distinct levels are recognized in the Jurassic history of foraminiferal evolution. At two of them (Pliensbachian–Toarcian and Bathonian–Callovian boundaries), the intensity of extinction of high-rank taxons was followed by the subsequent high renewal rate, whereas the third one, at the Aalenian–Bajocian boundary, is marked only by the high index of community renewal at the generic level. Indices of extinction in the Aalenian were low. This boundary was marked by the appearance of planktonic foraminifers, the group of which anticipated and determined the further devel-

Table 4. Main parameters of the dynamics of the generic composition of Jurassic foraminifers

Epoch	Age	Number of genera (G)	Quotient of renewal, % of A	Most important parameters of macroevolutionary transformations: Index of renewal, % of G	Index of appearance, % of G	Index of disappearance, % of G	Index of extinction, % of G	Extinction rate
J_3	Tithonian (Volgian)	137	26.4	16.7	8.0		13.1	1.9
	Kimmeridgian	133	31.0	15.0	6.8	14.2	6.8	
	Oxfordian	144	34.7	[23.6]	16.7	[21.5] →	[12.5]	1.7
J_2	Callovian	135	[40.4] →	[28.3] →	14.1 —	18.6	7.5	
	Bathonian	126	27.6	19.0	14.2	[22.2] →	[13.4] →	[7.4]
	Bajocian	117	[45.0] →	[42.7] →	[21.3]	12.8	1.7	
	Aalenian	73	36.0	15.0	12.3	6.8	2.7	
J_1	Toarcian	86	[38.3] —	19.7 —	15.1 —	27.9	1.1	
	Pliensbachian	94	31.3	[25.5]	[18.0]	[27.7] →	[8.5] →	[2.1]
	Sinemurian	77	26.5	[24.6]	14.2	10.3	6.5	
	Hettangian	60			[36.2]	3.3	1.7	

(A) Total number of genera in compared contiguous stages, (G) number of genera registered within the stage, [36.2] – extremum.

opment of the foraminiferal biota, and by the origin of five new families. The number of the latter is about two times greater than the corresponding average parameter for the Jurassic high-rank taxons.

Data considered above lead to the following conclusions.

(I) In dynamic aspect, the foraminiferal taxonomic diversity during the Jurassic was fluctuating, displaying a drastic fall at the Early–Middle Jurassic boundary (Toarcian–Aalenian), restoration in the Bajocian, subsequent growth until the Oxfordian, and stabilization during the terminal ages of the Jurassic. Stabilization was accompanied by the gradual decrease in origination rates of high-rank taxons.

(II) The following stages can be distinguished in the Jurassic foraminiferal evolution: (1) the Rhaetian–Lias crisis of biota accompanied by mass extinction of foraminifers; (2) the Early Jurassic period of the low taxonomic diversity in the Hettangian and Sinemurian, which sharply increased in the terminal Sinemurian, Pliensbachian, and Toarcian; (3) the Middle Jurassic stage that started with the crisis at the Toarcian–Aalenian boundary followed by the gradual diversity decrease in the Arctic and Boreal Atlantic zoochores during the Bathonian and by concurrent increase in the Tethyan zoochore; (4) the terminal Middle Jurassic bloom accompanied everywhere by an intense taxons radiation; (5) the Kimmeridgian–Tithonian epoch of stabilization.

(III) The most important event in the Jurassic foraminiferal evolution was the appearance of planktonic forms represented by four genera of the new order Globigerinida that happened at the beginning of the Middle Jurassic.

(IV) The Jurassic foraminiferal history includes some ages with the low background rate of genera origination (8–12 new genera per age) and other ages of their active radiation (from 17 to 30 genera per age). The highest rate of genera origination was characteristic of the Bajocian and Oxfordian ages.

(V) The Jurassic foraminiferal history also included migration periods of genera origination centers. The latter were connected with the Boreal-Atlantic zoochore from the Hettangian to Aalenian and with the Tethyan tropical areas from the Bajocian to Tithonian.

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Phytostratigraphy of Jurassic Coal-Bearing Formations of the Irkutsk Basin

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Abstract—The Jurassic stratotype and reference sections in the Irkutsk basin located along the Angara River and its tributaries are studied along with remains of fossil plants sampled in the bed-by-bed manner. Paleobotanical data are used to subdivide Jurassic sections into the phytostratigraphic units (beds with flora) and to substantiate their lateral correlation. Isolated sections are correlated between each other and their succession in the Jurassic sequence is defined. Based on the paleobotanical data, sedimentary formations of the Jurassic basin are correlated with the coeval deposits of southern West and Central Siberia. The age of the Cheremkhovo Formation, except for its upper member, is arbitrarily estimated to be of the terminal Early Jurassic (Toarcian) age, whereas the upper silty–sandy member is dated back to the initial Middle Jurassic, and the Prisayanskaya Formation is correlated with the Bajocian.

Key words: **paleoflora, plant assemblage, beds with flora, correlation, Jurassic, West Siberia.**

INTRODUCTION

The Jurassic coal-bearing formations widespread in the Irkutsk coal basin were known as early as in the last century, beginning from the earliest geological works of Chekanovskii (1874), Cherskii (1888), and Bogdanovich (1895). Korovin (1921) was first to propose subdivision of the coal-bearing sequence into four stratigraphic units (from the base upward): Cheremkhovo, Barkhatovo, Idan, and Sukhovskaya formations. In 1922, the same author (Korovin, 1922) united the Cheremkhovo and Barkhatovo formations into the single Cheremkhovo Formation enclosing four coal-bearing horizons and resting on the basal Zalari conglomerate horizon. The bed-by-bed description of outcrops of Jurassic deposits along the Angara River from Irkutsk to the Olonki Settlement was performed slightly later by Zhemchuzhnikov (1925) in the course of the detailed geological survey. Following Korovin, this author defined the conglomerate horizon at the base of the Jurassic sequence, but he united the Cheremkhovo, Barkhatovo, and Idan formations into the Lower Formation, whereas the Sukhovskaya Formation was termed as the Upper Formation. Zhemchuzhnikov was first to correlate isolated outcrops of Jurassic rocks in bank exposures along the Angara River and to compile an integral Jurassic section. Subdivision proposed by Korovin and Zhemchuzhnikov appeared to be most reliable. The subsequent stratigraphic–geological studies only specified their stratigraphic scheme. For instance, Deev (1957) proposed to consider the Zalari conglomerate horizon underlying the Jurassic sequence as the unit of the formation rank; he also suggested to rename the Lower Formation back to the Cheremkhovo Formation and the Upper Formation back to the Sukhovskaya Formation. In the same year, N.I. Fomin proposed to refer the upper part of the Jurassic sequence to the Kuda Formation, which, however, was considered by other researchers as representing a facies counterpart of the Prisayanskaya Formation (see Deev, 1957).

The most complete data on the structure and lithological–mineral peculiarities of the Jurassic coal-bearing sequence of the Irkutsk basin, as well as paleogeographic and paleofacies reconstructions are given in the monograph by Odintsova *et al.* (1967). The lateral correlation of Jurassic sequences proposed in the book was based mainly on the prospecting drilling, geological survey, and palynological data. Unfortunately, lists of floral macroremains were published without critical analysis, and all the available data on the Jurassic flora, beginning from determinations of O. Heer, were quoted. Such a consideration of paleobotanical materials did not contribute much the stratigraphy of host deposits. Therefore, the age of stratigraphic units accepted by mentioned authors was substantiated only by palynological data: the Cheremkhovo Formation was dated back to the Early Jurassic and the Prisayanskaya Formation to the Middle Jurassic.

First data on the Jurassic flora from the Irkutsk coal basin can be found in the monographs by Heer (1876, 1878, 1880), the Swiss scientist, who studied collections sampled by Chekanovskii in 1869. Heer studied plant remains from exposures located near Irkutsk (Mt. Kaya) and along the Angara River (the right-hand bank close to the mouth of the Ust'-Balei Creek and near the Tapka Ravine). He described more than 50 plant species. Comparing the Irkutsk flora with those from Europe, Heer assumed its Middle Jurassic age. The subsequent studies of the Irkutsk flora (Seward and

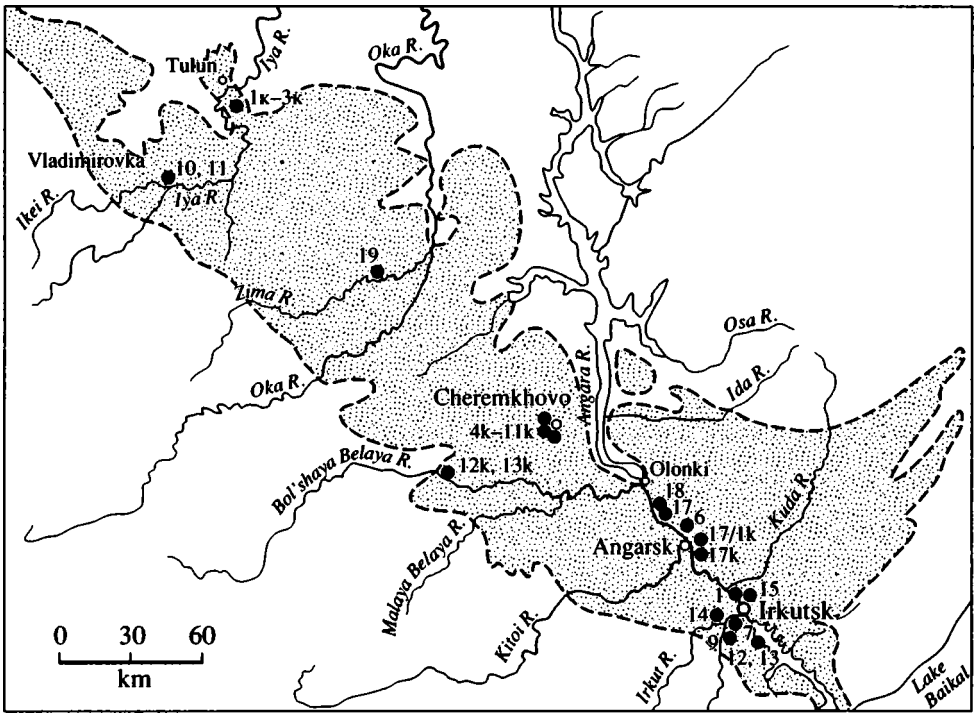

Fig. 1. Localities with fossil plant remains in the Irkutsk basin: (1k–3k) Azeya coal quarry; (12k, 13k) Belaya River, near the Kekurka Settlement; (17) right-hand bank of the Angara River, Volch'ya Ravine; (17/1k) right-hand bank of the Angara River, Idan Ravine; (1, 15) right-hand bank of the Angara River, Tapka Ravine; (6) right-hand bank of the Angara River, Ust'-Balei Creek; (7) Mt. Kaya, right-hand bank of the Irkut River; (9) Smolenshchina Village, Olkha River; (10, 11) Vladimirovka Village, Iya River; (12, 13) Irkutsk water store, Mel'nichnyi Ravine; (14) Maksimovshchina Village, Irkut River; (17) right-hand bank of the Angara River, opposite the Tel'ma railway station; (18) right-hand bank of the Angara River, Tolstyi Mys; (19) Basalaevka Village; Zima River. The dash line indicates boundaries of distribution of Jurassic deposits.

Thomas, 1911; Turutanova, 1920; Khakhlov, 1924, 1927) did not enlarge the previous taxonomic list, but some determinations of ferns were revised, and the age of previously defined formations was defined as the Middle Jurassic. Only Khakhlov (1924), who studied collections by Korovin and Zhemchuzhnikov, erroneously ascribed remains of coniferous stems of the *Pagiophyllum* and *Elatocladus* type to the genus *Sequoia* and mistook the leaf impression of the *Clathropteris*-type fern with reticular veining sampled from the section of the Irkut River for the leaf of the angiosperm genus *Cissites*. On this basis, Khakhlov suggested the Early Cretaceous age of the upper Sukhovskaya Formation, and the age of underlying formations was correspondingly estimated to be the Late Jurassic. The fallibility of these determinations was noted by Krishtofovich (1933), who considered the coal-bearing sequence of the Irkutsk basin as the Middle Jurassic in age. The monograph of Prinada (1962) dedicated to the Jurassic flora of the Irkutsk basin and Transbaikal region also did not clarified the problem of the age of coal-bearing deposits in question. Prinada revised some determinations by Heer, particularly those of ferns and some coniferous plants. He also specified determinations by Khakhlov for samples from the collection of Zhemchuzhnikov and other geologists and estimated age of the considered flora as the Middle Jurassic. However, despite the importance of work done by Prinada, who described more than 90 species, the latter cannot be used for the phytostratigraphic purposes. This is explained by the fact that most of the described plant remains lack the exact stratigraphic and even geographic allocation, and ginkgophytes, the dominant group of plant remains, were determined using only morphological features of leaves without regard for the peculiarities of their epidermis structure.

Ermolaev and Teslenko (1964) were first to propose the Early Jurassic age of the Cheremkhovo Formation based on the data on fossil macroflora. This assumption was inferred from the presence of the *Phlebopteris* leaf remains recovered by boreholes from the upper part of the formation and of *Clathropteris* remains cited by Khakhlov (1924). Unfortunately, paleobotanical materials of the mentioned paper also cannot be used for stratigraphic interpretation because of absence of exact geographic and stratigraphic allocation of plant remains. Therefore, the Interdepartmental stratigraphic meeting on the Mesozoic of Central Siberia held in Novosibirsk in 1978 accepted the stratigraphic scheme for the Irkutsk basin, where the age of previously

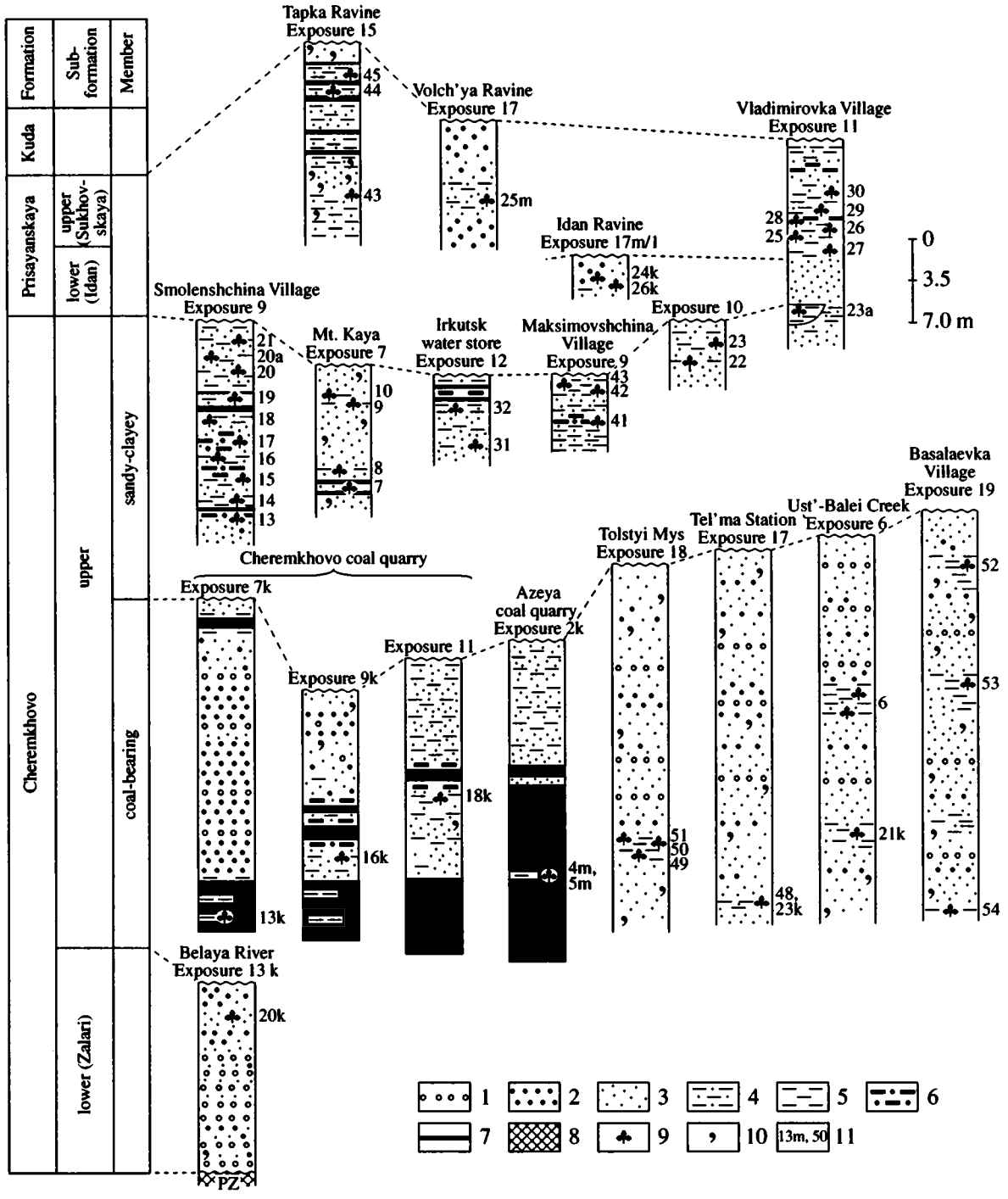

Fig. 2. Correlation of Jurassic sections in the Irkutsk basin: (1) conglomerate; (2) coarse-grained sandstone; (3) medium- to fine-grained sandstone; (4) siltstone; (5) argillite; (6) coaly siltstone; (7) coal; (8) underlying rocks; (9) remains of macroflora; (10) coalified plant remains; (11) samples with macroflora.

defined formations was substantiated by palynological data only (*Resheniya...*, 1981). According to resolutions of the meeting, the Cheremkhovo Formation was referred to the Early Jurassic (Pliensbachian–early Toarcian). The lower subformation of the Prisayanskaya Formation was also considered to be of the Early Jurassic (Toarcian) age, whereas the upper subformation was attributed to the Middle Jurassic (Aalenian).

In 1987–1992, we visited the stratotype and reference sections mentioned in the above-mentioned works (Heer, 1876; Korovin, 1922; Zhemchuzhnikov, 1925; Ermolaev and Teslenko, 1964) (Fig. 1). We described

these sections anew and sampled for the first time fossil macrofloras in the bed-by-bed manner. The thorough study on their taxonomic composition using the epidermal–cuticular method for determination of angiosperm leaves and revision of plant remains from the collection by Heer (Mt. Kaya, the Balei Creek mouth flowing in the Angara River), which are stored at the Geological Institute RAS, Moscow, significantly specified the taxonomic composition of the Jurassic flora from the Irkutsk basin. Characteristics of many previously known taxons such as *Ginkgo sibirica* Heer, *Sphenobaiera czekaniwskiana rigida* (Heer) Flor. *Phoenicopsis angustifolia* Heer, and *Czekanowskia rigida* Heer was specified and, in addition, new species of these genera were encountered. Solution of some nomenclature problems concerning the *Coniopteris* and *Raphaelia* ferns, and also some conifers changed previous views on their composition, distribution throughout the section, and age. The results of paleobotanical studies were used to correlate the Jurassic rocks in isolated exposures, to establish an integral succession of stratigraphic units, and to determine the age of the latter.

PHYTOSTRATIGRAPHY

The Jurassic coal-bearing sequence in the Irkutsk basin unconformably rests upon the Paleozoic strata and is subdivided into three stratigraphic units (from the base upward): the Cheremkhovo, Prisayanskaya, and Kuda formations (*Resheniya...*, 1981), The total thickness of the Jurassic sequence is about 500 m.

The Cheremkhovo Formation

Korovin (1922) was first to define the formation as corresponding to productive beds of the Cheremkhovo mine. At the meeting in 1978, the formation was accepted as a unit comprising the basal conglomerate horizon previously defined as the Zalari Horizon or Formation (Korovin, 1921, 1922; Deev, 1957) and the proper Cheremkhovo productive formation. The integral thickness of the formation is up to 250 m. The formation is subdivided into three members (*Resheniya...*, 1981). We follow Odintsova with co-authors (1967), who proposed a two-member subdivision of the Cheremkhovo formation into the lower Conglomerate (Zalari) and upper Productive subformations.

The lower Cheremkhovo (Zalari) Subformation was first defined as an autonomous stratigraphic unit (horizon) by Korovin (1922). It was later studied in detail by Otten (1934) and Deev (1957). The subformation is of a limited distribution range. The section located in the vicinity of the Zalari railway station represents its stratotype. The basal part of the subformation is composed of variegated conglomerates laterally replaced by sandy–conglomerate to sandy–clayey rock varieties. The subformation is about 100–150 m thick. The Zalari Subformation is observed in bank exposures along the middle courses of the Belaya River near the Kekurka Settlement (Fig. 1), where the Paleozoic rocks are overlain by the 15-m-thick sequence (Fig. 2), whose basal part corresponds to the conglomerate bed (1.5 m) higher replaced by a member of alternating conglomerate, sandstone, and siltstone with the following plant remains: *Equisetites* sp., *Cladophlebis haiburnensis* (Lindl. et Hutt.) Sew., *Cladophlebis* sp., *Raphaelia diamensis* Sew., *Czekanowskia rigida* Heer. The boundary between the lower and upper subformations of the Cheremkhovo Formation coincides with the base of the siltstone and sandstone member underlying the "Main" coal seam (Odintsova *et al.*, 1967).

The upper Cheremkhovo (Productive) Subformation about 200 m thick (according to the drilling data) is subdivided into two units: the lower coal-bearing and upper sandy–silty members with subordinate coal seams (Odintsova *et al.*, 1967).

The lower coal-bearing member encloses commercial coal seams being completely recovered in sections of the Cheremkhovo quarry near the eponymous town. The basal portion of the recovered section, which represents a stratotype section of the upper Cheremkhovo Subformation, corresponds to the "Main" coal seam 6 m thick and is overlain by the member of alternating sandstone, siltstone, and argillite grading upward into alternation of sandstone and siltstone with 3–4 coal seams (from 0.3 to 1.5 m thick). The latter are laterally spliced and replaced by sandstones and siltstones. The total thickness of the section exposed in the quarry is about 20 m. We sampled members of alternating rocks above the "Main" coal seams and layers located higher between coal seams (Fig. 2). More than 10 samples collected in the bed-by-bed manner yielded abundant plant impressions, often with excellently preserved phytoleims of angiosperm leaves.

First and, perhaps, the only data on the flora from the Cheremkhovo quarry can be found in the work of Seward and Thomas (1911), who described six plant species collected above the top of the "Main" coal seam. They determined *Equisetites* cf. *ferganensis* Sew., *Cladophlebis haiburnensis* (Lindl. et Hutt.) Sew., *Baiera* sp., *Ginkgo digitata* Brongn., *Phoenicopsis speciosa* Heer, and *Czekanowskia rigida* Heer. The assemblage dates the enclosing deposits back to the Middle Jurassic. We also sampled more than 20 species of taphoflora from the Cheremkhovo sequence exposed in the eponymous quarry. Most diverse amid them are representatives of Ginkgoales and Czekanowskiales (Table 1). The former group is dominated by *Ginkgo celebris* Kiritch., *G. sibirica* Heer, *Sphaenobaiera czekanowskiana* (Heer) Flor., *S. vigentis* Kiritch. et Batjaeva, *Sphenobaiera* sp. 1, and the latter one by *Czekanowskia baikalica* Kiritch. et Samyl., and *Cz. rigida* Heer. The fern group is less diverse. It includes transient forms of the *Cladophlebis haiburnensis* (Lindl. et Hutt.) Sew. and *Raphaelia diamensis* Sew. groups (Table 1). Abundant are also remains of *Equisetites lateralis* (Phill.) Phill. Coniferous taxons are rare in this locality.

Table 1. Distribution of the Cheremkhovo floral assemblage in various localities

Plant list \ Locality, sample	Cheremkhovo coal quarry								Ust'-Balei	Tolstyi Mys			Zima River, Basalaevka Village	Angara River, Tel'ma Station	Azeya quarry
	10k	11k	12k	13k	14k	15k	16k	17k, 18k		49	50	51			
* *Hepaticites* sp.									+						
Lycopodites tenerrimus Pryn.									+			+			
Phyllotheca sibirica Heer									+				sp.	cf.	
Equisetites asiaticus Pryn.						+									
E. lateralis (Phill) Phill.		sp.			+		+	+				+			
Equisetostachys sibiricus Heer									+	+	+			+	
Phlebopteris sp.		+													
Coniopteris murrayana (Brongn.) Brongn.									+					cf.	
Coniopteris sp.									+					+	
* *Cladophlebis argutula* (Heer) Font.									+						
C. haiburnensis (Lindl. et Hutt.) Brongn.		cf.		sp.		+	+	sp.	+		cf.	cf.		cf.	
C. cf. *vaccensis* Ward											+				
* *C. whitbiensis* Brongn.									+						
Lobifolia sp.					+					+					
Raphaelia diamensis Sew.								+	sp.					sp.	
Nilssonia iniqua Kiritch.		+					+								
** *Aegianthus sibiricus* (Heer) Krassil.									+						
** *Angarolepis odorata* Krassil. et Bugd.									+						
** *Cadmisega ephedroides* Krassil.									+						
** *Heerala antiqua* (Heer) Krassil.									+						
Baiera sp.									+				+		
Ginkgo celebris Kiritch.					+		+	+		+	+	+		+	+
G. concinna Heer									+						
G. sibirica Heer								+	+	+			sp.	+	
Sphenobaiera czekanowskiana (Heer) Flor.								+	+	+	+	+		+	
Sphenobaiera sp. 1				+				+		+				+	
Sphenobaiera sp. 2								+							
Sphenobaiera sp. 3													+		
S. vigentis Kiritch. et. Batjaeva	+				+		+	+		+		cf.		+	
Czekanowskia baikalica Kiritch. et Samyl.		+					+			+	+	+		+	
Cz. curta Kiritch. et Samyl.															+
Cz. jenissejensis Kiritch. et Samyl.										+					
Cz. rigida Heer	+			+	+		+	+	+	+		cf.		cf.	+
Cz. cf. *vilujensis* Kiritch. et Samyl.															+
Leptostrobus crassipes Heer									+						
L. laxiflora Heer									+	sp.		sp.	+		
L. microlepis Heer.	+														
Schizolepis cf. *moelleri* Sew.	+	sp.	sp.											sp.	
* *Schizolepidium gracile* Heer									+						
Pityophyllum nordenskioldii (Heer) Nath.		+												+	
Pityophyllum (*Florinia*?) sp.		+	+												
Pityospermum maakiana (Heer) Nath.									+						
Samaropsis rotundata Heer									+				+		
*** *Elatides ovalis* Heer									+	+				sp.	
*** *E. brantiana* Heer									+						
Elatocladus sp. 1															+
Elatocladus sp. 2															+
Sorosaccus sibiricus Heer									+	+				+	
* *Squama* sp.									+						
* *Carpolithes cintus* (Nath.)									+						
* *Antolithes schmidtianus* Heer									+						
Ixostrobus heeri Pryn.				+				+	+	sp.		sp.	+	sp.	
Pseudotorellia sp.					+			+							

Determinations after: *Prinada, 1962; **Krassilov and Bugdaeva, 1988; ***Doludenko and Kostina, 1987.

Taphofloras similar to the Cheremkhovo one were sampled in exposures along the right-hand banks of the Angara River from the Olonki Settlement to (upstream) the Idan Ravine, which were first visited by Zhemchuzhnikov in 1925. We described this portion of the section of the Cheremkhovo Formation with the bed-by-bed sampling in Exposure 17 at the Angara River right bank opposite the Tel'ma railway station and in the Exposure 8 at the Tolstyi Mys locality (Fig. 1). The section 30 m thick in total is composed here of yellow massive medium- to coarse-grained sandstones with conglomerate interbeds, coalified plant remains, and argillite–siltstone lentil saturated with plant and insect remains (Fig. 2).

Taphofloras from the Tel'ma and Tolstyi Mys localities are dominated by Ginkgoales and Czekanowskiales species belonging to genera identical to those from the Cheremkhovo section (Table 1): *Ginkgo* (*G. celebris* Kiritch., *G. sibirica* Heer), *Sphenobaiera* (*S. czekanowskiana* (Heer) Flor., *Sphenobaiera* sp. 1), and *Czekanowskia* (*Cz. baikalica* Kiritch. et Samyl., *Cz. rigida* Heer). Abundance and diversity of ferns remain to be insignificant in these rocks (Table 1). A similar list of characteristic species was compiled for the exposure at the Angara River near the Ust'-Balei Creek mouth, upstream of the Tolstyi Mys locality.

The Ust'-Balei exposure is represented by a yellowish sandstone block that is to 30 m thick in total and encloses two lentils of alternating siltstone, sandstone, and clay rapidly wedging out along the strike (Fig. 2). The lower lentil is composed of the coarser material and hosts rare plant remains mainly represented by leaves of *Czekanowskia, Sphaenobaiera*, and by fructifications of *Leptostrobus* and *Samaropsis*. The paleontological characteristic of the second silty–sandy lentil is more representative. Its roof layers composed of fine-grained light yellowish clays yielded abundant plant and insect remains. Plant remains first studied by Heer (1876, 1880), who described more than 50 species from this locality, originate most likely from the same lentil.

However, the recently performed revision of the Heer's collection significantly reduced his list of species. Doludenko and Rasskazova (1972) studied the leaf epidermis from type samples of some *Ginkgo, Sphenobaiera, Czekanowskia*, and *Phoenicopsis* species collected from the Jurassic sequences of the Irkutsk basin, the Ust'-Balei locality included, and first described by Heer. Species composition of the genus *Coniopteris* was specified by Kirichkova and Travina (1993) and taxonomic affinity of some gymnosperm fructifications by Doludenko and Kostina, (1987) and by Krassilov and Bugdaeva (1988). As a result of these studies, the list of plant species from the Ust'-Balei locality, with due account for additional sampling, includes 21 forms (Table 1). Nonetheless, this floral assemblage appeared to be not very representative in the stratigraphic aspect. According to dominant forms and characteristic species of genera *Ginkgo, Sphenobaiera*, and *Czekanowskia* (*G. sibirica* Heer, *S. czekanowskiana* (Heer) Flor., *Cz. rigida* Heer), the Ust'-Balei assemblage is most similar to the Cheremkhovo taphofloras. With respect of taphonomic importance, however, this locality remains unique, characterizing a paleoecological niche with peculiar phytological and enthomological communities.

We discovered similar burial of plant remains together with insects between the Basalaevka and Ignai settlements at the left-hand bank of the Zima River, a tributary of the Oka River. The exposed sequence of massive compact yellow and yellowish sandstones is up to 35 m thick here and encloses three thin siltstone interbeds with plant remains (Fig. 2). Similarly to the Ust'-Balei locality, taphofloras from these interbeds are mainly represented by gymnosperms, most abundant amid which are *Czekanowskia, Leptostrobus*, and *Sphenobaiera* genera, Equisetaceae species (*Phyllotheca sibirica* Heer), and some conifers (*Samaropsis rotundata* Heer, *Ixostrobus heeri* Pryn.) (Table 1).

In general, the Cheremkhovo taphofloras from all above-mentioned localities appeared to be sustained with respect of taxonomic composition (Table 1). The Equisetaceae family is represented by genera *Equisetites* and *Phyllotheca*. Ferns of the *Cladophlebis* and, less commonly, of *Raphaelia* genera are not numerous. The core of the Cheremkhovo taphofloras is composed of Ginkgoales (*Ginkgo sibirica* Heer, *G. celebris* Kiritch., *G.* spp., *Sphenobaiera vigentis* Kiritch. et Batjaeva, *S. czekanowskiana* (Heer) Flor., *Sphenobaiera* sp. 1, *Sphenobaiera* sp. 2), Czekanowskiales (*Czekanowskia baikalica* Kiritch. et Samyl., *Cz. rigida* Heer), and ancient conifers represented by fructifications (*Samaropsis*, Schizolepis, *Ixostrobus* strobiles, *Elatides, Stenomiscus*, and others). Taking into consideration the stable composition of the Cheremkhovo taphofloras, we consider it reasonable to define the corresponding deposits as Beds with *Ginkgo celebris, Sphenobaiera vigentis, Sphenobaiera* spp., and *Czekanowskia baikalica*. The lower 30-m-thick coal-bearing member of the upper Cheremkhovo Subformation, which is most completely exposed in the Cheremkhovo quarry, is proposed to be the reference section for this stratigraphic unit.

The upper, sandy–silty member of the upper Cheremkhovo Subformation is composed of alternating sandstones, siltstones, argillites, and thin coal seams changeable in lateral directions. These deposits are exposed in several outcrops (Fig. 1) near Irkutsk (vicinities of the Smolenshchina Village at the Olkha River, Maksimovshchina Village at the Irkut River, and Mt. Kaya), along the Angara (Irkutsk water store) and Iya rivers (near the Vladimirovka Village).

The most complete section of this portion of the Cheremkhovo Formation is exposed in the railway ditch at the Olkha River bank, a tributary of the Irkut River (opposite of the Smolenshchina Village). The basal part of the 20-m-thick section is composed of a

sandstone member replaced higher in the section by alternating siltstones, coaly siltstones, sandstones, clays, sandy clays, and thin (to 0.3 m) coal seams (Fig. 2). Siltstone and clayey interbeds yielded fairly diverse taphofloras (Table 2). Taphofloras from the Olkha River locality slightly differ in taxonomic composition from the Cheremkhovo ones. They are also dominated by gymnosperms including ginkgophytes. The role of the *Sphenobaiera* genus amid Ginkgoales sharply decreases: only one species is present instead of five species in the Cheremkhovo taphofloras. The *Ginkgo* genus retains its diversity, but is represented by different species (Table 2). Also present are genera *Erethmophyllum* and *Leptotoma*. Czekanowskiales are dominated by the genus *Phoenicopsis* that is missing from the Cheremkhovo taphofloras; the *Czekanowskia* genus is less common and represented by other species: *Cz. irkutensis* Kiritch. et Samyl., *Cz. vera* Kiritch. et Samyl. Ferns are More diverse in the Olkha taphofloras, particularly representatives of the *Raphaelia* and *Cladophlebis* genera. More common are leaf remnants of *Coniopteris*, in particular, *C. maakiana* (Heer) Pryn., and also *Hausmannia* leaves with the reticular veining. Conifers are less diverse than in the Cheremkhovo taphofloras being mainly represented by the *Pagiphyllum*-type stems with leaves and by isolated leaves of the *Pseudotorellia* and *Pityophyllum* genera.

Taphofloras similar to the Olkha ones were sampled in the Mt. Kaya section located at the mouth of the Kaya River, a tributary of the Olkha River. The Mt. Kaya section about 15 m thick is poorly exposed and consists of sandstone and sandy siltstone beds intercalated with two coal seems (Fig. 2). Sandy siltstone beds yielded plant remains dominated by leaves of *Coniopteris, Cladophlebis, Raphaelia*, and *Phoenicopsis* (Table 2).

Another section of the Olkha type is recovered near the Mel'nichnyi Ravine on the left-hand coast of the Irkut water store. The section is 15 m thick and mainly composed of alternating sandstones, siltstones, coaly siltstones, and coal seams (Fig. 2). The collection of plant remains sampled from this section includes 14 species, among which there are four *Phoenicopsis* species. Taphofloras from the Irkutsk water store locality show a higher diversity of conifers differing from those the Cheremkhovo taphofloras. These are *Elatocladus* and *Taxocladus (Florina?)* sp. Exposures of the upper member of the upper Cheremkhovo Subformation were also studied along the Iya River, in the vicinity of the Vladimirovka Village (Fig. 1). Directly near the village, the river-bank exposures exhibit the 4-m-thick sandstone sequence with rare siltstone interbeds enclosing plant remains (Fig. 2); 1 km downstream of Exposure 10, there are alternating sandstones, siltstones, and coal seams with plant remains (Exposure 11). Rare siltstone interbeds in Exposure 10 and sandstone beds enclosing the 1.5-m-thick siltstone lentil at the base of Exposure 11 (Fig. 2) yielded the plant assemblage similar in taxonomic composition to the Olkha taphoflora. The assemblage is dominated by the same species of the *Czekanowskia* and *Phoenicopsis* genera and by some representatives of conifers. Ferns in taphofloras from the Iya section are represented by the same species of the *Hausmannia* genus (Table 2). Higher in the section of Exposure 11, there is a sequence of yellowish to white (at the fresh surface) quartzose blocky sandstone that underlies the bedded rock member with another plant assemblage of different taxonomic composition. Most likely, Exposure 11 along the Iya River exhibits the boundary between the Cheremkhovo and Prisayanskaya formations (Fig. 2).

Thus, the upper part of the Cheremkhovo Formation (its upper member) is characterized in all sections by taphofloras of a rather stable taxonomic composition. We unite these taphofloras into the Olkha assemblage. Its dominant forms are *Coniopteris maakiana* (Heer) Pryn., *Cladophlebis, Raphaelia diamensis* Sew., diverse *Phoenicopsis* forms (5 species), *Czekanowskia irkutensis* Kiritch. et Samyl., and *Cz. vera* Kiritch. et Samyl. This provides grounds to classify the upper member of the upper Cheremkhovo Formation with the Olkha floral assemblage as Beds with *Coniopteris maakiana, Cladophlebis* spp., *Phoenicopsis dentata*, and *Phoenicopsis* spp. The section along the Olkha River, opposite of the Smolenshchina Village and section along the Iya River near the Vladimirovka Village are proposed to be the reference sections of this stratigraphic units. The thickness of beds with flora is 20–25 m.

The Prisayanskaya Formation

This formation conformably overlying the Cheremkhovo Formation is less widespread as compared with the latter being exposed mainly in the southeastern part of the Irkutsk basin (Odintsova *et al.*, 1967). Exposures of the formation are known along the right-hand bank of the Angara River from Irkutsk to the Idan Ravine. The total thickness of the unit is up to 130 m. The formation is subdivided into the lower Idan and upper Sukhovskaya subformations.

The lower Idan Subformation first defined by Zhemchuzhnikov (1925) in the Angara River basin as the Idan Horizon is widespread in the southeastern part of the Irkutsk basin (Odintsova *et al.*, 1967). In addition, the subformation is traceable near the Vladimirovka Village in the exposure at the right-hand bank of the Iya River (Fig. 2). The subformation is almost barren of plant remains and consists mainly of coarse-grained to gravelly sandstones with subordinate siltstone interbeds and. The thickness of the unit is up to 80 m; in the sections along the Angara River, it is about 45 m thick. The siltstone interbeds in the section near the Idan Ravine along the Angara River yielded abundant remains of *Coniopteris* sp. Indet., *Raphaelia* sp., *Sphenobaiera* sp. indet., *Czekanowskia vera* Kiritch. et Samyl., and *Podozamites* sp. indet.

The less widespread Sukhovskaya Subformation (Odintsova *et al.*, 1967) was first defined as the

Table 2. Distribution of the Olkha floral assemblage in various localities

Plant list \ Locality, sample	Mt. Kaya, Irkut River		Olkha River, opposite of the Smolenshchina Village				Irkut River, Maksimovshchina Village	Iya River, Vladimirovka Village		Irkutsk water store
	7, 8	9, 10	13, 14	15–18	19, 20, 20a	21		22, 23	23a	
Hepaticites sp.				+						
Lycopodites sp		+								
Phyllotheca sibirica Heer									+	
* *Clathropteris* sp.		+								
Hausmannia crenata (Nath.) Möeller					+				+	
* *Phlebopteris*(?) sp.					+					
Coniopteris irkutensis Pryn.			sp.	+			sp.		sp.	
C. maakiana (Heer) Pryn.	+	+			+	+				+
Cladophlebis bidentata Tur.-Ket.						+				
C. haiburnensis (Lindl. et Hutt.) Brongn.	sp.	+	+		sp.		sp.			
C. nebbensis (Brongn.) Nath.		+		+						
C. vaccensis Ward		+		+						
C. whitbiensis (Brongn.) Brongn.						+				+
Lobifolia sp.						+				
Raphaelia diamensis Sew.	+		+	+		+	+			+
Baiera chachlovii Kiritch. et Batjaeva						+				sp.
Eretmophyllum sp.				+						
Ginkgo sp. 1				+						+
Ginkgo sp. 2										+
G. tapkensis Dolud. et Rasskaz.								+		sp.
Leptotoma sibirica Kiritch. et Batjaeva						+				
Sphenobaiera sp. 4					+					sp.
Czekanowskia irkutensis Kiritch. et Samyl								+	+	
Cz. vera Kiritch. et Samyl.			+	+						
Phoenicopsis angustifolia Heer							+		+	+
Ph. dentata Pryn.	+		+	+						+
Ph. irkutensis Dolud. et Rasskaz.				+		+	+	+		+
Ph. markovitchiae Kiritch. et Schischk.	+									
Ph. samylinae Kiritch. et Moskv.							+			+
Leptostrobus sp.					+					
Brachyphyllum sp.					+					
Pagiophyllum peregrinum (Lindl. et H.) Sew.				+					+	
Pityophyllum sp.					+			+		+
Pityospermum sp.						+				+
Elatocladus sp.										+
Taxocladus sp.										+
Pseudotorellia sp.				+					+	

Determinations after: * Prinada, 1962.

Table 3. Distribution of the Tapka floral assemblage in various localities

Plant list \ Locality, sample	Right-hand bank of the Angara River			Iya River, Vladimirovka Village			
	Kashtak Ravine	Volch'ya Ravine	Tapka Ravine	Sample 25, 26	Sample 28, 29	Sample 30	Sample 27
* *Neocalamites pinitoides* (Chach.) Pryn.		+	+				
Phyllotheca sibirica Heer			+				
** *Osmunda sibirica* Kostina et Travina			+				
Hausmannia crenata (Nath.) Möeller			+				
Coniopteris irkutensis Pryn.	+		+				
C. maakiana (Heer) Pryn.			+	+			cf.
* *Cladophlebis* ex gr. *crispata* Racib.		+					
C. nebbensis (Brong.) Nath.	sp.		+				
Lobifolia lobifolia (Phill.) Rassakaz. et E. Lebed.			+				
*** *Raphaelia tapkensis* (Heer) Pryn. emend. Kostina	sp.	cf.	+				
Anomozamites lindleyanus Schimp.			+				
Cycadolepis sp.			+				
* *Podozamites* sp.		+					
Baiera chachlovii Kiritch. et Batyaeva							+
Baiera. sp. 1							+
Ginkgo heeri Dolud. et Rasskaz.		+	sp.				
Ginkgo sp. 3		+					
G. tapkensis Dolud. et Rasskaz.			+	+	+		
Sphenobaiera sp. 5	sp.	+					+
Czekavowskia cf. *ferganensis* Kiritch. et Samyl.			+				
Cz. irkutensis Kiritch. et Samyl.				+			
Cz. rigia Heer				+	+		
Cz. vera Kiritch. et Samyl.				+		+	+
* *Phoenicopsis dentata* Pryn.		+					
Phoenicopsis irkutensis Dolud. et Rasskaz.			+	+		+	
Leptostrobus crassipes Heer						+	
L. laxiflora Heer		sp.	+	+		sp.	
Schizolepis sp.				+			
Pityocladus sp.			+				
Pityophyllum (*Florinia*?) sp.			+				+
Pityospermym cf. *gracile* Tur.-Ket.				+			
Samaropsis rotundata Heer				+			
Elatites ovalis Heer			+				
Elatocladus manchurica (Yok.) Yabe			+				
Taxocladus ketovae Tesl.		sp.	+				
* *T. sibirica* (Chach.) Tesl.			+				
Sorosaccus sp.		+					
Stenomiscus sp.						+	
Ixostrobus sp.						+	
Pseudotorellia sp.					+		

Determinations after: * Prinada, 1962; ** Kirichkova *et al.*, 1999; *** Doludenko and Kostina, 1997.

Sukhovskaya Formation by Korovin (1921), and later as the upper subformation by Zhemchuzhnikov (1925), in the bank outcrops extending along the Angara River from Irkutsk to the Idan Ravine. The subformation is composed of alternating fine- to coarse-grained sandstones, siltstones, coaly siltstones, and coal seems. Sandy rock varieties are preponderant. The total thickness of the unit is up to 50 m. Studying the subformation at the right-hand bank of the Angara River from the vicinity of Irkutsk to Idan Ravine, we traced a sequence of compact sandstones bearing limy concretions and intercalated with three siltstone and sandy siltstone interbeds of variable thickness. The most complete section was studied near the Tapka Ravine (Fig. 2), where the 20-m-thick sandstone sequence encloses siltstone and coaly siltstone interbeds and thin coal seams. In outcrops located 1.5 km downstream, the sequence encloses a single siltstone member with coal seams; another section near the Kashtak and Volch'ya ravines (Sukhovskoi Cape) located 20 km downstream is also composed mostly of sandstone with a single member of sandy siltstones (Fig. 2). Siltstone interbeds from these exposures enclose plant remains, most diverse of which are those in the Tapka Ravine locality (Table 3). The Tapka taphofloras, as well as those from the Kashtak and Volch'ya ravines localities (Fig. 2) differ from the Olkha taphofloras in the composition of dominant plant groups despite the apparent affinity. The fern group reveals a sharply reduced share of representatives of the *Cladophlebis* and *Raphaelia* genera and the appearance of the *Osmunda* genus. The *Ginkgo* genus is represented by the same species as in the Olkha taphofloras. Representatives of the *Phoenicopsis* genus almost completely disappear among Czekanowskiales. Conifers become significantly more diverse owing to representatives of the *Elatocladus* and *Taxocladus* genera and large leaves of the *Florina* type (Table 3).

Similar taphofloras were sampled from the bedded sequence overlying the Idan Subformation of the Prisayanskaya Formation in the Iya River section near the Vladimirovka Village (Fig. 2). These taphofloras differ from the Tapka ones in a higher abundance of *Ginkgo tapkensis* Dolud. et Rasskaz., *Czekanowskia* species, particularly *Cz. vera* Kiritch. et Samyl., conifers, and presumable representatives of the *Florina* genus. Based on consistency of taxonomic composition of the Tapka taphofloras, we propose to define the upper half of the Prisayanskaya Formation (Sukhovskaya Subformation) as Beds with *Osmunda sibirica, Raphaelia tapkensis, Czekanowskia vera,* and *Taxocladus.* The section along the Angara River near the Tapka Ravine is proposed to be the reference section for this unit. The thickness is about 15 m.

The Jurassic section in the Irkutsk basin is crowned by the Kuda Formation (Table 4). The formation was defined by N.I. Fomin in 1946, but some researchers considered it as a facies analogue of the upper part of the Prisayanskaya Formation (Deev, 1957). In line with resolutions of the Interdepartmental meeting, the unit is accepted in the formation rank (*Resheniya...*, 1981). The unit is characterized by a limited distribution in the northwestern areas of the basin (Odintsova *et al.*, 1967), where it is represented by basal conglomerates laterally replaced by sandstones and, higher in the section, by siltstones and argillites, light gray almost white with yellowish shading. The total thickness is up to 150 m. The rocks are almost barren of plant remains, except for poorly preserved leaves of *Coniopteris* sp, *Cladophlebis* sp., and *Phoenicopsis* sp. (Odintsova *et al.*, 1967).

AGE SUBSTANTIATION

The bed-by-bed paleobotanical characteristic of the reference and stratotype sections and revision of taxonomic composition of fossil plants provide the first opportunity to substantiate the stratigraphic subdivision of the Jurassic formations in the Irkutsk basin on the basis of taphofloras and to estimate age of different stratigraphic units. The analysis of changes in taxonomic composition of taphofloras throughout the section revealed the stratigraphic allocation of particular floral assemblages and evolutionary trends of the Irkutsk paleoflora. As a typical representative of the Siberian paleofloral province, the Irkutsk paleoflora shows the same evolutionary trends, which are typical of the Early–Middle Jurassic floras of southern West Siberia. These trends are particularly pronounced in composition of the ginkgophytes (Ginkgoales and Czekanowskiales) dominant amid the Siberian paleofloras and ferns. Although these groups are represented by different species in separate areas of Siberia, for instance in the Kuznetsk and Kansk basins, their distribution and relationships throughout the section, i.e., secular variations, reveal common tendencies. Thus, determining the age of defined phytostratigraphic assemblages from the Irkutsk basin, one should perform their correlation with due account for data on flora genesis in the adjacent regions, in particular, in the Kansk and Kuznetsk basins.

The most complete Jurassic section of southern West and Central Siberia is that in the Kuznetsk coal-bearing basin, where Jurassic deposits are subdivided into lithostratigraphic and stratigraphic units having the detailed paleobotanical characteristics (Kirichkova *et al.*, 1992). Investigations of paleoflora in the Kuznetsk basin revealed that the most significant changes in composition of the plant assemblages occurred in the middle of the Osinniki Formation, where the Bungarap paleoflora is replaced by the Chernyi Etap plant assemblage. Above this level, the *Coniopteris* genus represented by 7 fern species and cicadophyte *Nilssonia* genus are important components of taphofloras from the upper Osinniki Subformation against the background of diverse ferns of the *Cladophlebis* genus. Nonetheless, the most pronounced transformation characterizes ginkgophytes dominant in the Kuznetsk paleoflora. In the lower Osinniki taphofloras, this group includes mainly the *Sphenobaiera* and *Czekanowskia*

Table 4. Correlation of Jurassic sections of the Irkutsk, Kansk, and Kuznetsk coal-bearing basins based on the paleobotanical data

genera, whereas the upper Osinniki and Chernyi Etap taphofloras are characterized by abundant *Ginkgo, Phoenicopsis*, and *Leptotoma* representatives, but the *Sphenobaiera* and *Czekanowskia* genera are less diverse here and represented by other species. Similar changes are recorded across the boundary between the Pereyaslovka-Ilan and Kamala formations in the sections of the Kansk coal-bearing basin (Kirichkova and Travina, 1990) and in the upper part of the upper Khudoseya Subformation of the Tomsk region in the West Siberian basin (Kirichkova and Travina, 1995). Taking into consideration the concept of Vakhrameev on development of the Jurassic floras at the Early–Middle Jurassic boundary (Vakhrameev, 1982; Vakhrameev *et al.*, 1970), these changes, along with regional peculiarities of fossil floras in the Siberian paleofloral province, were used to substantiate the Early–Middle Jurassic boundary in the continental section of southern West Siberia (Kirichkova *et al.*, 1992). Similar differences in composition between the Cheremkhovo and Olkha taphofloras are evident in the coal-bearing section of the Irkutsk basin. The Cheremkhovo floral assemblage from the lower part of the upper Cheremkhovo Subformation (its coal-bearing member) reveals proportions between characteristic forms similar to those in the Kamzas assemblage from the lower Osinniki Subformation of the Kuznetsk basin. In addition, it includes the same guide forms: *Ginkgo sibirica* Heer, *Sphenobaiera vigentis* Kiritch. et Batjaeva, *Czekanowskia baikalica* Kiritch. et Samyl. At the same time, the Olkha taphofloras characterizing the upper part (sandy–silty member) of the upper Cheremkhovo Subformation shows the same changes in taxonomic composition as the Chernyi Etap taphofloras from the upper Osinniki Subformation of the Kuznetsk basin. Although the Olkha taphofloras bear apparent endemic features, they characterize the level of grooving abundance of *Coniopteris* forms in the assemblage still retaining the high species diversity of the *Cladophlebis* genus. Moreover, the parallel changes are characteristic of ginkgophytes, the dominant plant group. At the Olkha level, this group includes different *Ginkgo* species and less diverse *Sphenobaiera* genus also represented by other species. Simultaneously, this is the first occurrence level of the *Leptotoma* genus, and, what is more important, the *Phoenicopsis* genus becomes most widespread here. Consequently, we used these changes to substantiate the Lower–Middle Jurassic boundary in the integral continental section of the Irkutsk basin, like it was done in the Kuznetsk and Kansk coal basins. We consider this boundary as corresponding to the base of the upper sandy–silty member of the upper Cheremkhovo Subformation of the Cheremkhovo Formation (Table 4). Accordingly, the age of beds with the Cheremkhovo floral assemblage is estimated to be the terminal Early Jurassic (presumably, Toarcian); the Olkha floral assemblage appears to correspond in the development level to the Chernyi Etap assemblage from the Kuznetsk basin and can be dated back to the initial Middle Jurassic (arbitrarily, Aalenian), and the beds with the Tapka floral assemblage are attributed to the Bajocian (Table 4).

CONCLUSION

(1) The bed-by-bed study of the Irkutsk paleoflora allowed us to specify its taxonomic composition, to determine the degree of its endemism, and to define particular taphofloras and their exact stratigraphic range. The analysis of changes in taphoflora composition through the section revealed certain trend in the development of the Irkutsk paleoflora, which is similar to that characterizing the Siberian paleoflora in general.

(2) The characteristic phytostratigraphic assemblages (Cheremkhovo, Olkha, Tapka) are distinguished. They are used to define beds with flora. The succession of beds in the section is established and their lateral distribution is elucidated.

(3) The first correlation scheme is proposed for isolated sections, whose position in the integral Jurassic section of the basin is established using the paleobotanical data.

(4) Position of the Lower–Middle Jurassic boundary in the study region is substantiated by changes in composition of the Irkutsk paleoflora, which are similar to those previously established in taphofloras from the Kuznetsk and Kansk basins and probably characterizing the Jurassic flora of entire southern Siberia. This boundary in the Jurassic section of the Irkutsk basin corresponds to the base of the upper sandy–silty member of the upper Cheremkhovo Subformation. Accordingly, the remaining part of the Cheremkhovo Formation is estimated to be of the terminal Early Jurassic age (the Toarcian and probably a part of the Pliensbachian); the Prisayanskaya formation is arbitrarily dated back to the Bajocian.

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New Data on the Natural Residence Environment of Primitive Man and Age of the Sungir' Archeological Monument (Vladimir Oblast)

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Abstract—Based on the new 14C dates, it is estimated that the age of the Sungir' archeological monument ranges from nearly 29 to 20 ka. The history of the Sungir' man includes two stages of his permanent residence at the site and several substages of intermittent visits to this territory. Stages of the permanent residence of the Sungir' community correspond to periods of 28800–28350 and 27800–25500 years ago, whereas substages of episodic visits, to periods of 28120–28000 and, approximately, 25000–20000 years ago. The palynological data reveal a complex dynamics of the landscape–climatic changes at the end of the Valdaian, when the permanent residence stages of the primitive Sungir' man corresponded to periods of climatic warming.

Key words: **middle Valdaian, ^{14}C age, palynological histograms, Sungir'.**

The available views on the age of the Sungir' archeological monument and natural environments of primitive man are quite variable (Bader and Gromov, 1963; Tseitlin, 1965; Gromov, 1966; Sukachev *et al.*, 1966; Bader, 1978; Lavrushin and Spiridonova, 1998). This is related to the fact that there are only few radiocarbon age determinations, which were questioned by some researchers, whereas palynological data are almost completely missing.

Lavrushin and Spiridonova (1998) were first to publish complete palynological characteristics of the Sungir' section, according to which its largest portion, the cultural–soil layer included, was referred to the Valdaian.

Simultaneously, the thorough geological investigations revealed at least two members of buried soils intensely deformed by gravitational slope processes and enclosing remnants of the Sungir' material culture. Palynological studies showed that the primitive man site was initially surrounded by the spruce forest area with meadows overgrown with herbage and cereals, and also by marshy landscapes in river valleys. Later, during the second period of buried soil accumulation, the spruce forest was replaced by a mixed formation of dominant pine and subordinate spruce and birch. Open landscapes were occupied, as previously, by herbage meadows and bogs, areas of which probably increased.

NEW RADIOCARBON DATES FOR THE OCCUPATION LAYER

New 14C dates were obtained using bone fragments of large mammals extracted from the occupation layer during archeological studies of the monument under consideration (table).

The dates indicate that the late Paleolithic community of primitive peoples first appeared at the site in question approximately 29 ka ago. The latest signs of man residence in this area are dated back to 20360 years ago. Thus, the primitive man visited this site during the period of almost 8500 years. Most of the dates

Radiocarbon age of bone remains of larger mammals (the Sungir' archeological monument)

Sample number	Radiocarbon age	Animal bone fragments
GIN 9585	20360 ± 900	Mammoth
GIN 8998	23600 ± 500	"
GIN 8995	26300 ± 260	"
GIN 9034	26300 ± 300	Horse
GIN 9030	26600 ± 300	Mammoth
GIN 9591	27000 ± 320	"
GIN 9027	27200 ± 400	"
GIN 9586	27200 ± 500	"
GIN 9033	27400 ± 400	Horse
GIN 9588	27800 ± 600	Mammoth
GIN 9029	28000 ± 300	"
GIN 8999	28120 ± 170	"
GIN 8996	28130 ± 370	"
GIN 9032	28350 ± 200	"
GIN 9028	28800 ± 240	"

obtained fall into the time interval of 27–28 ka ago. This can be explained by the fact that the primitive peoples resided at the Sungir' site most likely in this period, although the residence history could be discontinuous to a certain extent.

The younger dates most likely indicate episodic visits of ancient hunters to the site. Moreover, judging from the obtained dates, these visits could be separated by the period of 1000–1500 years. It is noteworthy as well that there are no dates corresponding to the intervals of 20360–22000, 22000–23500, 23500–24500, 24500–25500, and 25500–26300 years.

The new radiocarbon dates listed in the table are obtained using, as a rule, a single bone fragment of a large mammal. The only exception is Sample 9034 representing several fragments of horse bones. This average date was excluded from the further analysis.

Thus, the new dates obtained substantially contribute to our knowledge of the Sungir' monument age and outline a long-lasting period, when primitive peoples visited the study area.

PALYNOLOGICAL RESULTS

New palynological data on the Sungir' site were published recently. They shed light on stratigraphy and paleogeography of two occupation layers confined to different-age buried soils, and characterize general environmental changes during the larger part of the Valdaian, including the Dunaevo (Bryansk) interstade, as well as cooling and onset of the late Valdaian warming period. An additional study on peculiarities and consistency of palynological spectra composition throughout the exposed part of the occupation layer were performed in the excavation of 1993. In this work, we present new materials on the spores–pollen assemblages, which are dated by the radiocarbon method (Fig. 1). Spores and pollen were extracted from the mineral residue after demineralization of bone fragments and collagen cleaning.

The obtained palynological results are of a certain interest with respect of methodology, because they allow a more detailed subdivision of the intensely deformed occupation layer, thus outlining periods of the primitive man residence at the site. It is also important that every sample contains abundant well-preserved microfossils, sometimes accompanied by common herbaceous pollen.

Let us consider palynological assemblages in their time succession. The first (I) spore–pollen assemblage (Samples 9032 and 9028 dated back to 28350 ± 200 and 28800 ± 240 years ago, respectively) is dominated by the arboreal pollen (80%) including that of *Pinus, Picea*, and *Betula* sect. *Albae* (pollen of the latter two forms amounts to 10–15%); the *Alder* pollen occurs as single grains. The herbaceous pollen is equally represented by cereals, sedges, and wormwood. The assemblage also includes pollen of Asteraceae, *Plantago,* and Cichonaceae. Predominant amid sporogenous taxa are *Bryales* and Polypodiaceae accompanied by rare *Sphagnum* and *Botrychium.*

The second (II) spore–pollen assemblage (Samples 8999 and 8996 dated back to 28120 ± 170 and 28130 ± 370 years ago, respectively) differs from the previous one in the arboreal pollen composition. Similarly to the first assemblage, this pollen is abundant, but dominant is spruce pollen, whereas pine, birch, and alder pollen is less common. The composition of herbaceous formation retains its main features. Preponderant cereal and sedge taxa associate with subordinate Chenopodiaceae and some herbage forms. Similarly to the previous assemblage, spores mostly belong to *Bryales* and Polypodiaceae.

The established principal differences in composition of arboreal pollen in the first and second assemblages imply that the period from 28350 to 28130 years ago was marked by substantial environmental changes, because the pine forest was replaced by the spruce one presumably in response to the climatic warming. It is not inconceivable that climatic changes also influenced the primitive man life, although this is not registered by archeological data.

The third (III) spore-pollen assemblage (Sample 9029 dated back to 28000 ± 300 years ago) also bears specific features suggesting the onset of cooling. For instance, the share of *Betula* sect. *Albae* increases to 23% despite the fact that the arboreal formation is dominated, as earlier, by the pine pollen. The share of spruce and alder pollen does not exceed 2–3%. In addition, the content of the herbaceous and frutescent pollen also increases, especially the abundance of sedge and wormwood pollen. The share of the Chenopodiaceae pollen is insignificant.

The fourth (IV) spore–pollen assemblage (Samples 9030, 9591, 9027, 9586, 9033, 9588) corresponds to the time interval lasting from 27800 to 26600 years ago (approximately 1200 years). This relatively long interval is characterized by the less persistent composition of palynological spectra. The content of dominant arboreal representatives in spectra is variable, and the total content of the arboreal pollen is relatively low (55–43%). The content of the birch pollen is persistently high, but the abundance peaks of spruce and pine pollen show alternation. The minimal content of the arboreal pollen in spectra is noted in Sample 9027, the radiocarbon age of which is 27200 ± 400 years. The herbaceous formation is characterized by the increased share of sedge and decreased content of cereals and Chenopodiaceae forms. The composition of spores appears to be more conservative. The noted changes in palynological spectra reflect corresponding dynamics of landscape transformation apparently related to short-term climatic variations. Moreover, warming-to-cooling transitions were sufficiently rapid. In general, this period was characterized by the most unstable climatic regime peculiar of the progressive warming phase. The vegetation cover

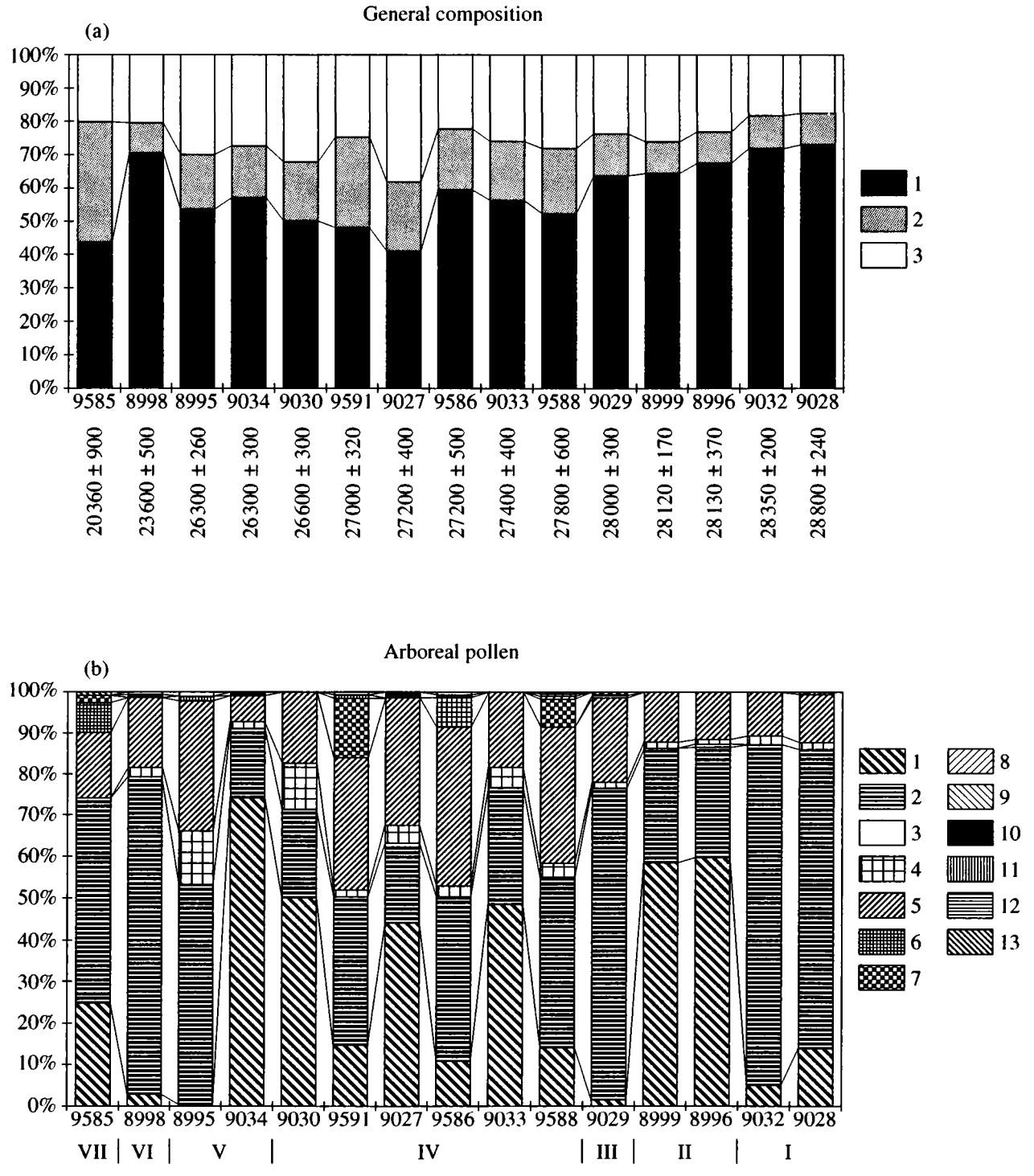

Fig. 1. Palynological histograms characterizing mineral residue extracted from bone remains of larger mammals. (a) General composition: (1) spores, (2) herbaceous pollen, (3) arboreal pollen. (b) Arboreal pollen: (1) Hippophae, (2) *Tilia*, (3) *Ulmus*, (4) *Quercus*, (5) *Corylus*, (6) *Salix*, (7) *Betula nana*, (8) *Betula* sect. *Fruticosae*, (9) *Betula* sect. *Albae*, (10) *Alnus*, (11) *Pinus Hapl.*, (12) *Pinus*, (13) *Picea*. (c) Herbaceous pollen: (1) Poaceae, (2) Cyperaceae, (3) Chenopodiaceae, (4) *Artemisia*, (5) *Typha*, (6) Jridaceae, (7) *Fagopirum*, (8) *Polygonum*, (9) *Rumex*, (10) *Plantago*, (11) Ranunculaceae, (12) Brassicaceae, (13) Dipsacaceae, (14) Apiaceae, (15) Cichoriaceae, (16) Asteraceae, (17) *Centaurea*. (d) Spores: (1) Polypodiaceae, (2) *Lycopodium selado*, (3) *Botrychium*, (4) *Equisetum*, (5) *Sphagnum*, (6) Bryae. Roman numbers designate palynological assemblages.

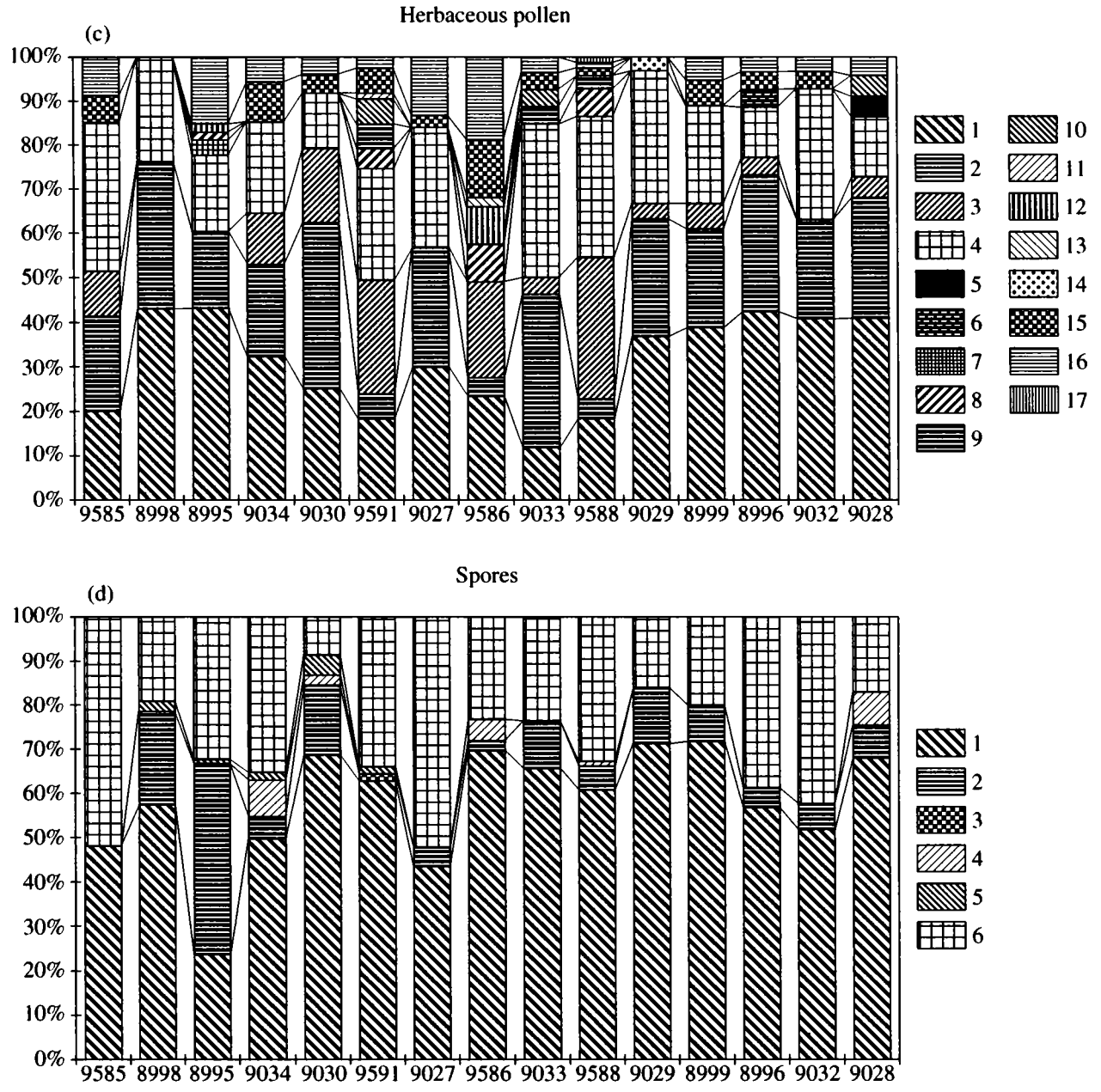

Fig. 1. (Contd.)

experienced sharp changes. During the coldest episode 27200 ± 400 years ago, the forest density in the study area was minimal. This episode probably coincided with the phase of thinned or patchy forests composed of birch, pine, and spruce. As evident from the general pollen spectra composition and from the increased share of sedge pollen, the area was substantially boggy. During the warmer phases, forests were thicker, having an increased share of pine and spruce.

According to palynological characteristics of this assemblage, at least three phases of environmental changes can be distinguished during the corresponding period. They are reflected in composition of arboreal and herbaceous pollen and, partly, of spores. The radiocarbon dates show that duration of every phase was about 400 years.

The fifth (V) spore–pollen assemblage (Sample 8995, the ^{14}C age 26300 ± 260 years) is characterized by appearance of pollen of broad-leaved plants in addition to the prevalence of pine and birch pollen and increased share of alder pollen amid the arboreal formation. The herbaceous taxa are dominated by cereals, sedges and wormwoods, which commonly associate with herbage pollen. Preponderant amid sporogenous

taxa are *Sphagnum*, Polypodiaceae, and *Bryales*. The considered assemblage indicates warming. Moreover, it should be noted that the short period that lasted from 26600 to 26300 years ago (about 300 years long) was probably marked by the northward migration of distribution areas of representatives of the European temperate forests.

The sixth (VI) spore–pollen assemblage (Sample 8998, the ^{14}C age 23600 years) reveals the preponderance of pine pollen in the arboreal group, where share of spruce, alder, and birch is insignificant. Prevailing among herbaceous pollen is that of cereals, sedges, and wormwoods. The assemblage also reflects sufficiently mild climatic conditions. It should be emphasized, however, that this warming phase was separated from the previous one by a period of about 2700 years. According to the available data, this period corresponded to climatic cooling, which is registered in some sections of European Russia (Spiridonova, 1983).

The seventh (VII) spore–pollen assemblage (Sample 9585, the ^{14}C age 20300 ± 900 years) differs from all previous assemblages in the substantially lesser content of arboreal pollen and in the increased proportion of herbaceous and frutescent plants. The assemblage suggests more open landscapes. In addition to arborescent and frutescent birch varieties, the palynological spectrum includes pollen of *Betula nana* L. and single well-preserved grain of sea buckthorn. The studied site was most likely surrounded by patchy forests formed by pine and birch. The forest areas adjacent to the valley side probably had an increased share of spruce. Open landscapes were occupied, as previously, by meadows of variable composition and by bogs, whose role noticeably increased. These landscape characteristics well correspond to the initial stage of the last glacial maximum (LGM).

DISCUSSION AND CONCLUSION

The new data allow us to reconsider age of the archeological monument, the natural habitat environment of primitive man, and duration of his residence at the site located in the right-hand side of the Sungir' Creek.

As was unanimously noted by many researchers, the occupation layer of the monument is intensely deformed. This hampers, to a significant extent, the detailed stratigraphic subdivision of the layer using traditional geological methods. The above-mentioned radiocarbon dates and palynological data allow us to specify the age of the monument, habitat environment of primitive man, and stratigraphy of the occupation layer.

Earlier, several radiocarbon dates were obtained for the occupation layer (Bader, 1978), and some of them were considered as rejuvenated. Nonetheless, others were taken for sufficiently reliable, at least by some researchers; they are as follows: 22500 years for the wood coal from the base of Burial 1, 24500 years for the fragment of the reindeer bone, and 27700 years for the fragment of the bison bone. These dates in combination with those listed in the table suggest that primitive man visited the Sungir' area during the period 8500 years long at least. However, distribution of dates throughout the section reveals two periods of the primitive man residence in the area under discussion. The first period corresponds to dates clustered within the interval of 25500–28800, which indicate the permanent residence of primitive man in this area. Burials at the Sungir' residence sites of primitive man, which are discovered by archeologists, probably correspond to this period.

The second period bracketed by dates of 20300 and 25500 years can be characterized as a stage of the episodic visits of primitive man to the area under consideration. This assumption is inferred from several available dates separated from each other by substantial time intervals of about 1000–2500 years long. It is important to note that Burial 1 should be referred to this period. It is not inconceivable as well that Burial 2 is at the same level.

The first period can be probably subdivided, with a sufficient confidence, into several substages separated from each other by the short-term time intervals, when primitive man did not live at the Sungir' site. These substages are detectable based on both the distribution of radiocarbon dates throughout the section and on the results of palynological study. At least three substages can be outlined. The oldest of them corresponds to the first spores–pollen assemblage and is restricted by dates of 28350 and 28800 years. The middle stage characterizing most likely the short-term (episodic) residence of primitive man at the site coincides with the distribution range of spore–pollen assemblages II and III dated back to 28120 and 28000 years ago, respectively. The youngest substage is dated within the time interval of 27800–25500 years ago; it comprises the spore–pollen assemblages IV and V marking, similarly to the oldest substage, the permanent residence of primitive man at the site. Unfortunately, these substages are not substantiated by the detailed analysis of archeological materials, thus being rather arbitrary.

The discussed palynological materials suggest that the Sungir' community lived in different landscape settings. As shown in Figure 2, the Sungir' primitive man occupied the area surrounded by forests of the pine, birch, spruce and, even, thin-forest types. Thus, the Sungir' community survived different paleoclimatic regimes with fluctuations of the interstadial–stadial type. Nonetheless, it can be assumed that the first defined stage corresponded, in general, to the gradual warming in contrast to the second cooling stage.

The data obtained imply an intricate dynamics of climate during the phase of the progressing warming (Fig. 2), when the short-term climatic variations were about 400 years long. The most important peculiarity of

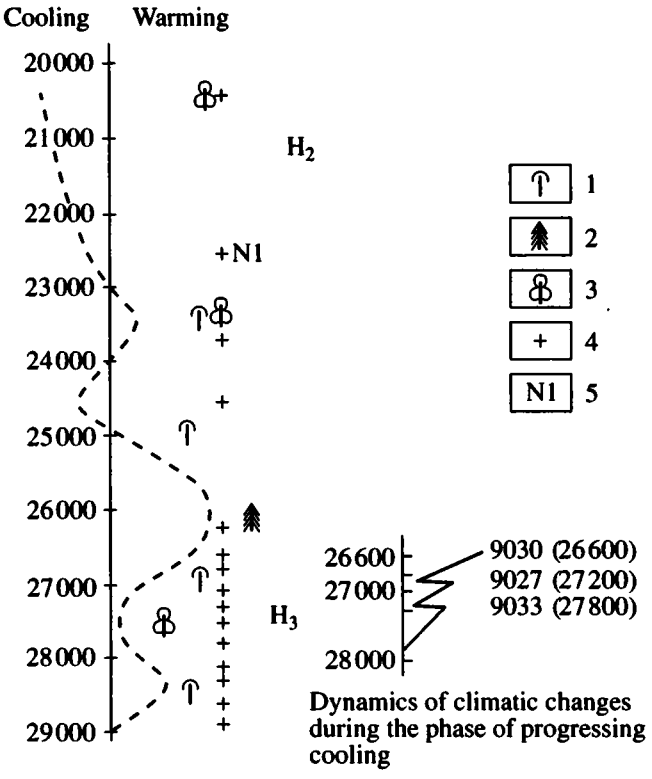

Fig. 2. The paleoclimatic curve inferred for the interval of 29–20 ka ago: (1) pine, (2) spruce, and (3) birch forests; (4) radiocarbon dates; (5) location of Burial 1; (H_2, H_3) Heinrich events.

these climatic fluctuations is an alternation of relatively long-lasting warming phases and short-term cooling phases. As a result, the curve demonstrating dynamics of climatic changes looks like a saw-shaped pattern with asymmetrical teeth.

In conclusion, we should emphasize the following.

(1) New radiocarbon dates for the Sungir' archeological monument are obtained using bone remains of larger mammals.

(2) Two stages and several substages are detectable in the Sungir' primitive man history.

(3) The Sungir' community of primitive peoples resided in sufficiently variable landscape–climatic settings.

(4) The inferable climatic changes were characterized by an intricate dynamics that is consistent with well-known climatic events of the interstadial–stadial type, which occurred in the Russian plain at the end of the middle Valdaian. Some of these events can be quite reliably correlated with the Heinrich events (Bond *et al.*, 1993), in particular, with the events H_2 and H_3, the radiocarbon ages of which coincide with our dates obtained for some climatic cooling episodes.

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Reviewer M.N. Alekseev

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Tamara Fedorovna Vozzhennikova (1914–2000)

Tamara Fedorovna Vozzhennikova, a prominent scientist, one of the founders of the national school of paleontologists, passed away on February 23, 2000.

Vozzhennikova was born on February 7, 1914 in the Village of Nizhnyaya Vokhminka in the Vyatka region. Being sixteen, she entered first the worker's department and then the Biological Faculty of the Leningrad State University, which she graduated from in the late 1930s. She always conceived of herself as a disciple of Professor K.M. Deryugin, an eminent zoologist, hydrobiologist, initiator and editor of a many-volumed issue *Studies of Seas of the USSR*. He organized several scientific marine stations, and Vozzhennikova worked at those stations before the Great Patriotic War. The war has halted her research work for several years. She has worked at an aircraft factory in Novosibirsk for 5 years participating in assembling of battle planes. She returned back to biology only in 1945, when she became a postgraduate student at the Chair of Hydrobiology of the Leningrad University. She defended her PhD dissertation *Algal Flora of Mountain Rivers in Tajikistan* in 1949. At the same year, she joined the Medical–Biological Institute, the West-Siberian Branch of the USSR Academy of Sciences.

In 1950, Vozzhennikova began to study an abundant group of Mesozoic and Cenozoic fossil algae in the Russian and West-Siberian plates heading the palynological team at the Institute of Mining Geology, the West-Siberian Branch of the USSR Academy of Sciences. She is rightfully regarded to be a veteran of the Siberian academic science. Vozzhennikova joined the Institute of Geology and Geophysics, Siberian Branch of the USSR Academy of Sciences, from the moment of setting up the branch, i.e. in 1958, and continued working there until the mid-1990s, when she moved to Moscow to her daughter's family because of illness.

Vozzhennikova was a splendid expert of all groups of ancient algal flora; she studied them in a close contact with other well-known algologists of this country: A.P. Jouse, I.A. Kiselev, A.V. Topachevskii, and others. She was at the head of the Laboratory of Paleophytology, one of the largest in Siberia, and acted as Chairwomen of the Paleoalgological Section of the USSR Palynological Commission. In 1965, the First All-Union Conference of Paleoalgologists was held in Novosibirsk under her supervision. As a member of the Organizing Committee, she contributed much to the preparation and execution of the Third All-Union (1964) and the Third International (1971) Paleontological conferences in Akademgorodok (Novosibirsk), the headquarter of Siberian Branch of the USSR Academy of Sciences.

As mentioned above, the main line of investigations of Vozzhennikova was the study of Mesozoic and Cenozoic organic-walled microphytoplankton, the most important group of fossils widely used nowadays for elaborating the detailed stratigraphy of oil-and-gas basins in both boreal and tropical zones, and also for solving paleogeographic and paleohydrological problems. She went into studying the microphytoplankton when only a small group of specialists in this country and abroad was involved in proper investigations. The problem to what kingdom (animal or plant) dinoflagellates, the key representatives of the microphytoplankton, belong has not yet been solved, though K.G. Ehrenberg, a German naturalist, revealed them as early as in the 1830s. Problems of the taxonomy and morphology of dinocysts remained especially entangled.

The issue of her first major monograph *Introduction to the Study of Fossil Algae Dinophyceae* (1965), which was highly appraised by "fathers" of the dinocyst stratigraphy J. Deflandre and A. Éisenack, was preceded by the publication of three important works: *Paleoalgologic Characteristics of Meso-Cenozoic Rocks of Western Siberia* (1960), *On the Taxonomy of Fossil Dinophyceae* (1961), and the review of pyrophytic algae, which was prepared for *Treatise of Paleontology* (1963). The chronological review of the investigation results on fossil and recent dinoflagellates beginning from their discovery to the 1960s (the monograph of 1965) represents a reference book of paramount importance. The works listed above, along with the second monograph issued in 1967, formed the basis for her doctoral dissertation defended in the same year in Novosibirsk. Studies carried out by Vozzhennikova were of great importance for developing paleoalgology in the world. In addition to the taxonomy and morphology of fossil dinoflagellates, a wide range of problems was touched upon in her works. Vozzhennikova was the first to focus the attention on paleoecological aspects, evolution, and generic relationships of dinoflagellates, because she showed, on the basis of materials from Russia, the perfect possibility of using this fossil group in stratigraphy. The monograph *Fossil Dinophyceae from Jurassic, Cretaceous and Paleogene rocks of the USSR* issued in 1967 was especially valued for the contribution to the dinocyst taxonomy and for description of many new taxons which later became common in

stratigraphy. Also of great interest is her next monograph *Dinocysts and their Stratigraphic Implications* (1979) devoted to hystrichosphaerids, classification of this group, and to detailed description of dinocysts. These new issues proved her authority as an outstanding specialist in algology widely known in the world.

Interest in organic-walled microphytoplankton, which arouse in the world by the early 1970s due to its wide practical use, was keen in this country as well. As a recognized leader, Vozzhennikova became head of the national school developing this line of paleoalgology. Her first disciples A.S. Andreeva-Grigorovich, K.E. Aristova, V.I. Il'ina, I.A. Kul'kova, É.S. Oleinik, V.N. Fedorova-Shakhmundes successfully studied dinocysts and prasinophytes during that period in different regions of the USSR (Crimea–Caucasus region, East European Platform, Siberian and Turan plates, Tajik Depression); they defended dissertations as well. For the purpose of studying the phylogeny and ecology of dinoflagellates, she took part repeatedly in cruises of the research vessel *Vityaz* to the Pacific Ocean and visited Sevastopol to study recent microphytoplankton of the Black Sea at the Institute of Limnology.

A vivid speech of Vozzhennikova at the conference of the national group of the IGCP Project 174 "Geological Events of the Terminal Eocene," which was held in Novosibirsk in the mid-1980s and visited by more than a hundred biostratigraphers from different regions of the former USSR, substantially increased the number of scientists involved in the study of the Mesozoic and Cenozoic organic-walled microphytoplankton. In 1987, 1989 and 1991, there were held in Moscow, under her supervision, the workshops for young specialists at the Geological Institute of the Russian Academy of Sciences, where she shared her experience in fundamentals of morphology and taxonomy of dinocysts. The number of her disciples increased noticeably. Teams oriented to investigation of this fossil group were founded in scientific centers and production enterprises of Kiev, Lvov, Minsk, Moscow, Leningrad, Yekaterinburg, Erevan, Alma-Ata, Tyumen, Okha, and other cities. The Paleoalgological Section has become deeply involved in this research. This resulted first of all in a greater number of papers on fossil dinocysts presented by Russian scientists at the international and All-Union palynological conferences. Vozzhennikova carried out active scientific investigations until the early 1990s when she revised, together with Judy Lentin (Canada), some *Vetzeliella* representatives. Their work "The Fossil Dinoflagellate Cysts *Kisselovia* emend. and *Charlesdowniea* gen. nov. Lentin et Vozzhennikova" was published in the *Review of the Palaeobotany and Palynology* and widely used in practice by paleoalgologists, as that group of cavate cysts turned out to be most important for the Paleogene zonation based on dinocysts. At the same time, Vozzhennikova published an article devoted to dinocysts found in Devonian rocks and opened a new page in the geological history of these microfossils. The last monograph by Vozzennikova "Fossil Dinoflagellates from Jurassic, Cretaceous and Paleogene Deposits of the USSR," which was written in collaboration with J. Lentin (Canada), was published abroad in 1993.

Investigations of dinocysts in Russia were not ceased, when Vozzhennikova stepped out of the research and organizational work because of her illness. Workshops have been revived in recent years (1997, 1999). More than 10 papers presented at the All-Union Palynological Conference (1999) were devoted to Mesozoic and Cenozoic dinocysts. The activity of regional teams has increased greatly, especially in Novosibirsk where Vozzhennikova's disciples Il'ina, Lebedeva, Kul'kova and their own young followers successfully work, elaborating the detailed stratigraphic zonations based on dinocysts from the Boreal Realm.

In the album devoted to pioneers in studying the organic-walled microphytopankton, which was edited on occasion of the International Dinocyst Conference held in Trondheim in 1999, Vozzhennikova was figured prominently among founders of the dinocyst stratigraphy.

Vozzhennikova left a fond memory of a splendid scientist and teacher in hearts of colleagues and disciples.

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